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METASOMATIC GRANITIZATION OF BATHOLITHIC DIMENSIONS[1]

PETER MISCH

PART I

ABSTRACT. In the Nanga Parbat area (Northwest Himalayas) a thick group of presumably pre-Cambrian argillites with thin calcareous and basic bands is overlaid by Cretaceous-Eocene volcanics, and both are bordered by extensive Eocene norite masses. These rocks have undergone regional metamorphism during Early Tertiary orogeny. Synkinematic metamorphism of the argillites progresses from slates and phyllites through micaschists, biotite-paragneisses, kyanite-schists and -paragneisses to sillimanite-paragneisses. The metamorphic zones in the argillites are matched by those in the calcareous bands.

Synkinematic granitization of the potash-predominant variety begins within the kyanite zone with microperthite-porphyroblast-schists passing to potash-rich augengneisses of granitic appearance which contain relict kyanite in their groundmass. Lit-par-lit replacement—as against mechanical injection—along active foliation planes makes banded gneisses. In the interior of the Nanga Parbat massif, granitization becomes more complete, and coarse-grained granitic gneiss forms. Here only small quantities of argillite altered to sillimanite-schist have escaped transformation into granitic rock. The calcareous bands are not granitized, and are preserved as conformable intercalations in the gneissose granite which has replaced the argillites. Metamorphism and granitization progress both along and across the strike of the altered sediments. There is complete structural continuity from the weakly metamorphosed sedimentary areas to the intensely granitized region. Also the type of tectonic deformation is the same throughout these areas, and is that characteristic of the deformation of solid rocks during orogenic metamorphism. The granitized body as mapped measures 25 by 60 miles.

The chemical nature of the granitizing agent is discussed. The metamorphic isograds (approximately vertical in this region) are shown to be independent of depth, and a function of differential introduction of heat from below. It is stressed that degree of metamorphism and granitization are systematically linked, and convection of heat by the granitizing solutions is held responsible for high grade regional metamorphisms in the upper part of the crust.

Introduction

THE idea that granitic rocks may form by metamorphic processes, is not new. It dates back to the French school

[1] Read during the April 1948 annual meeting of the Cordilleran Section of the Geological Society of America at Pasadena.

of the last century, and to the subsequent classical contributions by J. J. Sederholm (1907, 1923, 1926, 1934). However, only during the last dozen years or so has the problem of granitization received ever-increasing general attention. There is no need for giving a historical review here, since several recent publications have done so, among which the Geological Society's granite memoir is outstanding. It is significant that various geologists, working in widely separated parts of the world, have arrived at more or less similar conclusions, quite independently of each other, and often without previous knowledge of published reports of earlier students of the problem. It appears that geology has at present reached a stage in which there is inherent a tendency to re-examine fundamental conceptions and theories sacred by tradition, and to substitute new ideas where critical evaluation of the ever-increasing body of observed facts so demands.

This is especially true of the theory of the origin of granitic rocks—one of the basic problems in geology, considering the predominance of such rocks in the crust, and the role they play in the tectonic history of mobile belts. That granitic rocks can form by metamorphic transformation of pre-existing solid rocks on a comparatively small scale is today accepted by the great majority of geologists, since this process has been conclusively demonstrated by a number of authors, usually with an emphasis on small-scale features. Most geologists seem to feel that such processes are, as a whole, subordinate, and probably represent contact phenomena occurring in the marginal zones of igneous granitic batholiths. However, some of the workers in this field believe that metamorphic granitization is a process of much more fundamental and general significance. To prove this point, it has to be demonstrated that large masses of granitic rocks have as a whole, and not only marginally, formed by metamorphic transformation of solid rocks. The present paper attempts to contribute to this problem by describing some areas in which large masses of granitic rocks can be shown to have a metamorphic origin throughout.

The material presented was collected in the high mountains of Asia, but this is merely accidental; it appears to me that the areas described are not in any way extraordinary, but typical of what is found all over the world. The petrogenetic

aspect of the Himalayan gneisses described below is identical with that of many gneisses ranging from pre-Cambrian to Mesozoic ages which I have seen in this country, in China and in Europe, just as the "small granodioritic blocks formed by additive metamorphism" described by G. E. Goodspeed (1937) from Northeast Oregon are duplicated in the Okanogan country in Eastern Washington, in the Indus Valley in the Northwest Himalayas, or near the Yangtse Valley in Northwest Yunnan.

The material presented below is arranged according to whether granitization was synkinematic, i. e. took place under conditions of active orogenic deformation, or whether it was postkinematic, i.e. took place under static conditions when orogeny had ceased. Then some cases are briefly described which might serve to illustrate the relationships between syn- and postkinematic granitization. At the end of each part, it is attempted to draw certain general conclusions, presuming that the material presented is typical.

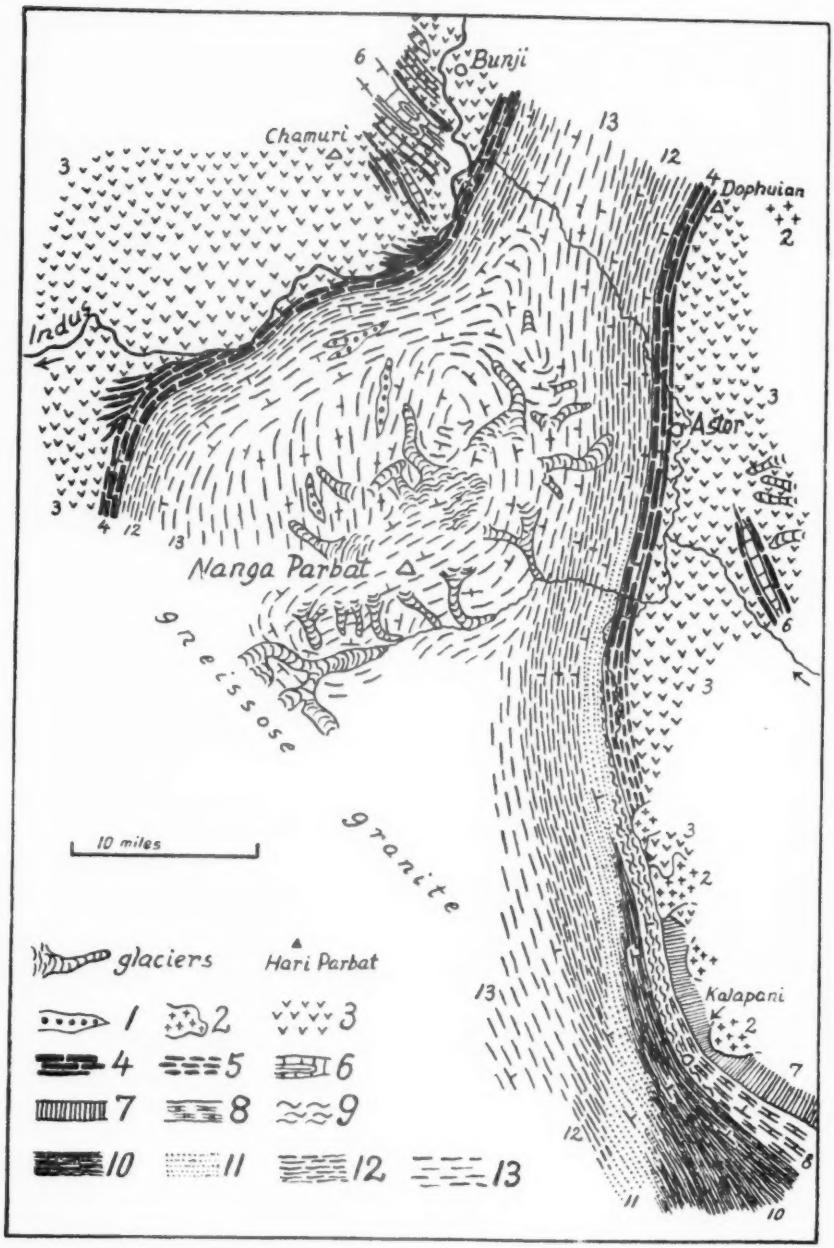
FIRST PART

Synkinematic granitization in Nanga Parbat area Northwest Himalayas

The Nanga Parbat area[2] in the crystalline inner zone of the Northwest Himalayas offers an unusually clear example of synkinematic granitization on a batholithic scale. Passages from metamorphic sediments of lowest to such of highest grades, and from chemically unaltered to intensely granitized rocks, are magnificently exposed in continuous sections for dozens of miles and over a vertical range of 23,000 feet. No structures or processes other or later than the main process of progressive alteration are superimposed, so that this process is revealed in grand simplicity.

Nanga Parbat Mountain rises north of the Great Himalayan Range, close to the upper Indus. The Nanga Parbat area shows a bending to north of the more westerly main trend of

[2] The writer did four months of field work in this area during Willy Merkl's 1934 Nanga Parbat expedition. The petrographic study has occupied a number of years since. A report on field observations has been presented (Misch, 1935) but, apart from a short abstract (Misch, 1936a), publication of the petrographic results has been delayed by emigration to China and the outbreak of the Sino-Japanese war, and the ensuing lack of adequate facilities in the interior of China.


Bunji
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Indus
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the Himalayas. This area forms a part of the eastern limb of the great syntaxial loop of the Northwest Himalayas which D. N. Wadia (1931) has described in detail in the sedimentary zones farther south.

(1) MAIN ROCK UNITS OF NANGA PARBAT AREA

The least metamorphosed bedded rocks are found in the southeast of the area shown in text figure 1, and farther south. Here occurs a *group of black phyllites and slates* with subordinate thin limestone bands, and occasional thin layers of quartzite and altered greenstone. The original thickness is very great and tectonic accumulation by intense folding is added. The resulting thickness exceeds six miles in the southeastern part of the region shown in figure 1, and south of this region the phyllite-slate group forms a still much thicker section. In the southeastern part of the region, this group is succeeded on the east by variegated tuffaceous and calcareous slates which in their upper part are interbedded with tuff breccias. Still higher in the section, there appear basic and also more acidic lava flows which in their lower part are intercalated with the tuffaceous rocks. This *volcanic formation* has been shown by D. N. Wadia (1937) to be *Cretaceous-Eocene*, on the basis of fossils discovered further southeast. Wadia

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Text Fig. 1. Geologic sketch map of Nanga Parbat region, Northwest Himalayas. After P. Misch (1935), fig. 8.

1: tourmaline-granite; postorogenic
2: quartz-diorite and granodiorite; post-norite
3: norite, hypersthene-diorite, etc.; widely altered to metanorite and metadiorite; Eocene
4: mesozonal amphibolite } synorogenic-metamorphic norite etc.
5: epizonal greenschist }
6: marble and schist zones in norite, metanorite, etc.
7: basaltic and more acidic lavas }
8: variegated tuff breccias and tuffaceous-calcareous argillites } Cretaceous-Eocene volcanic formation
9: tuffaceous argillaceous rocks (8) epizonally metamorphic }
10: Salkhala group; dark slates and phyllites with subordinate thin limestone and greenstone bands
11: non-granitized schists; chiefly mesozonally metamorphic Salkhalas
12: mesozonal migmatitic gneisses (chiefly augengneisses and banded gneisses); incompletely granitized Salkhalas
13: migmatitic granitic gneiss; granitized Salkhalas
(11)-(13) contain intercalations of schists, marbles, and amphibolites; mesozonal in (11) and (12), katazonal in (13)
} metamorphism and granitization synorogenic

(1932) assigns the underlying black slate group to the *pre-Cambrian Salkhala series* of the Western Himalayas. According to this author, the Salkhalas are everywhere separated by an angular unconformity from the Late Algonkian Dogra slate and all younger rocks. In the area here discussed, no angular relationship is visible in the field between the black slate group and the volcanic formation, their directions of foliation being parallel. In the tuffaceous rocks the foliation is essentially parallel to the bedding. In the black slates and phyllites, no bedding is visible in the field. Under the microscope, some of these rocks show intensely folded fine-bedding traversed by the foliation; isoclinal folds have their limbs parallel to the foliation and become drawn-out and finally obliterated by slip along the foliation planes. Any angular unconformity which may originally have existed is thus masked by the foliation.[3] The foliation is later than the volcanic formation and has originated during the late or post-Eocene main Himalayan orogeny.[4]

The black slates and phyllites and their limestone and greenstone bands pass gradually into *mesozonal crystalline schists* toward the west and north, that is, both across and along the strike. After progressive metamorphism has reached a mineral facies corresponding to the lower part of the mesozone, potash-feldspar porphyroblasts appear in the dark-colored schists[5] (chiefly biotite-paragneisses[6]) of argillaceous

[3] According to Wadia, such intense minor folding is characteristic of the Salkhalas though later structures may be superimposed.

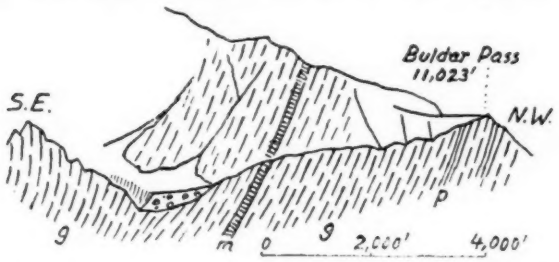
[4] The Himalayan folding is often called "Late Tertiary." However, the folding of the geosynclinal main range is Early Tertiary. Only the foldings of the successive foredeeps, marginal thrusting of the main range over the fore-deep rocks, and some local disturbances within the main range are Late Cenozoic (cf. Misch, 1936b, with references).

[5] The term *schist*, without a prefix, is here used to indicate any schistose rock derived from an argillaceous sediment, and having reached a higher metamorphic grade than phyllite. "Schist" thus includes micaschist, paragneiss and related rocks. "Crystalline schist" is a comprehensive term including all foliated metamorphic rocks. Schistose rocks differing in their composition from schists of argillaceous derivation, are indicated by a qualifying prefix or adjective, e.g. "chlorite-schist," "greenschist" (epidote-albite-chlorite-schist), "actinolite-schist," "calcareous schist," "quartzite-schist," etc.

[6] The term *paragneiss* is here used for crystalline schists having as main constituents quartz, mica and feldspar, and derived from sediment. Most paragneisses are fine-grained, rich in biotite and dark-colored. They may be thinly foliated, but are mostly somewhat more compact. They differ from micaschists in having feldspar as one of their main constituents, and mostly in being less thinly foliated.

derivation which thus grade into *mesozonal migmatitic augen-gneisses and banded gneisses*, both across and along the strike. The mesozonal rocks form a continuous belt on the eastern side of the Nanga Parbat gneiss massif, and a similar belt occurs in the northwest on the opposite side of the massif.

In the mesozonal belts, granitization remains incomplete, and material of purely sedimentary derivation is still recognizable everywhere. Toward the *interior of the gneiss massif*,



Text Fig. 2. Section and view south of Buldar Pass, northeast of Nanga Parbat.

g: coarse-grained granitic gneiss
m: white marble
p: fine-grained dark-colored biotite-schist (paragneiss)

granitization reaches a much higher degree. Most of the argillite-derived fine-grained dark-colored paragneiss becomes more coarsely recrystallized and feldspar-enriched, and thus loses its distinctive character. Here the mesozonal migmatitic gneisses grade into *coarse-grained light-colored granitic gneisses or gneissose granites*. These rocks form the bulk of the interior of the gneiss massif. Layers of dark-colored argillite-derived katazonal paragneiss, and of banded gneiss containing clearly discernible paragneiss laminae, are quantitatively very subordinate, but are intercalated throughout the interior of the massif. The granitic gneiss also contains layers of kata-amphibolites and of marbles with katazonal lime-silicates. These rocks may be associated with some paragneiss, but also occur directly intercalated in granitic gneiss. Particularly striking are a few places where pure white marble forms regular bands of small thickness but great longitudinal extent immediately in the coarse-grained gneissose granite (cf. text figure 2).

Field observations suggest, and microscopic structures con-

firm, that the progressive regional metamorphism is ***synkinematic***. The same holds true of the process of granitization though in the interior of the gneiss massif there are indications that crystallization in general and granitization in particular continued after deformation ceased.

In addition to the black slate group (Wadia's Salkhalas), the *Cretaceous-Eocene volcanic formation* occurring in the southeast of the area, *has undergone regional metamorphism.* In the extreme southeast of the area, alteration does not go beyond the slaty stage, and disappears on the east in the upper members of the volcanic formation. Toward the north, however, the metamorphism of the volcanic formation increases along the strike. Chlorite-schists, calc-phyllites, intensely laminated tuff-breccias with chloritic matrix, etc. represent the epizone. Farther north, hornblende-garbenschiefer and calc-micaschists, often rich in epodite-clinozoisite minerals, show the facies of the uppermost mesozone. Still farther north, some normal mesozonal amphibolites probably prolong this stratigraphic unit.

Another group of rocks in the Nanga Parbat area consists of very extensive masses of *norite and hypersthene-diorite* with local dunite. These rocks border the gneiss massif both on the east and northwest. The eastern norite area is on the south in contact with the predominantly basic lavas of the Cretaceous-Eocene volcanic formation. The norite is contemporaneous with or slightly younger than the lavas. That both are genetically related has already been pointed out by D. N. Wadia (1932).

Most of the noritic rocks have been more or less metamorphosed. There are all passages to statically recrystallized metanorites and metadiorites with crystalloblastic hornblende, epidote, etc. As the borders of the gneiss massif are approached, static recrystallization gradually gives way to kinematic metamorphism,[7] and *mesozonal amphibolites* are formed. The mineral association remains the same, except for the appearance of garnet in the deformed rocks. At the borders of the gneiss massif, the norite-amphibolites are

[7] The noritic series, its relation to large areas of included metamorphic Salkhalas where alternations and gradational contacts are observed, the later metamorphism of the noritic rocks, and the relations between their static and kinematic types of metamorphism, will be dealt with in another paper.

intimately intercalated and interfolded with the paragneisses and migmatitic gneisses. Both rock units, e. g. the norite bodies and the gneiss massif, have here participated in the same process of tectonic deformation and of crystallization accompanied by some granitization. This process of synkinematic metamorphism took place during the Early Tertiary main Himalayan orogeny, not a very long time after the original norite bodies had formed. At contacts with granitized bands, the amphibolites tend to be locally biotized.

Whereas the metamorphism of the norites is generally mesozonal, it becomes epizonal (chlorite - epidote - albite - schists, etc.) along the western border of the eastern norite body, where this border approaches the epizonal southeastern portion of the Nanga Parbat area.

Directionless granitic rocks occur in several parts of the area, but remain subordinate. Small irregular patches and pockets of quartz-dioritic and granodioritic rocks showing gradational contacts and porphyroblastic textures are common in directionless or weakly schistose metanorites. Some larger bodies of granodiorite occur in the noritic rocks and the lavas of the volcanic series. Within the gneiss massif, dikes and larger bodies of postorogenic white tourmaline-granite occur at several places, but the largest attain a size of only a few miles.

The *boundaries* between the metamorphic rock series described are all steep. The *metamorphic isograds* are steeply inclined planes in the whole area. For the most part, they are nearly perpendicular, the dip being either normal or slightly overturned.

(II) PROGRESSIVE METAMORPHISM OF SEDIMENTS IN NANGA PARBAT AREA

Argillite series.—This is the problem confronting us: do the granitic gneisses which form the bulk of the Nanga Parbat massif, owe their origin to intrusion of granitic magma, or to metasomatic[8] granitization? I shall try to give evidence for the latter process, e. g. for a formation of the granitic

[8] The term "metasomatic" is here used in its widest sense, i.e. to indicate any alterations of rocks during which chemical changes are brought about by an introduction of new substances, which process may or may not involve a simultaneous removal of other substances.

TABLE 1. MINERAL RANGES* IN METAMORPHIC SEDIMENTS OF

temperature zones	EPIZONE "highest" / "lowest"	MESO "upper"
ARGILLITES (in ARGILLACEOUS QUARTZITES same minerals, with larger proportion of quartz)	SERICITE SERICITE or MUSCOVITE — always present CHLORITE frequently present BIOTITE sometimes pres. (mostly sericitiform) Mn-rich garnet sometimes present ALBITE ALBITE or ALB-OLIGOCL. — often present QUARTZ	MUS usually much BIOTITE ALMANDITE POTASH FL OLIG widely present alwa
CALCAREOUS ARGILLITES and ARGILLACEOUS LIMESTONES	SERICITE SER. or MUSC. — always present CHLORITE frequently present BIOTITE often present (often sericitiform) Mn-rich garnet sometimes present ALBITE ALB-OLIGOCL or ALBITE — often present EPIDOTE-CLINOZOISITE frequently present CALCITE (DOLOMITE) always present QUARTZ in Mg-rich rocks (dol. shales) ACTINOLITIC HORNBLENDE sometimes present	MUS frequently present BIO mostly present POTASH FL ALMANDITE OLIG widely present EPIDOTE-CLINOZ ITE & ZOISITE frequently presen CALCITE nearly alwa in Mg-rich roc HORNBLE sometimes to
types of synkinem. granitized rocks		

* rutile (in argillites), sphene (in impure calcareous rocks), iron oxides, zircon, a

OF NANGA PARBAT REGION, NORTHWEST HIMALAYAS

"...r" / "lower" ZONE	KATAZONE
C / OVITE (often less, sometimes absent)	MUSCOVITE less or absent ("conditionally stable")
...h / nearly always present	BIOTITE always present
KYANITE frequently present	SILLIMANITE often present
...TE commonly present	ALMANDITE often present
... FELDSPAR frequently present	ORTHOCLASE nearly always present
O / CLASE	OLIGOCLASE nearly always present (more calcic plagioclase only in calcareous argillites)
...nt / nearly always present	
...ays present	QUARTZ present except in rare cases
C / OVITE	DIOPSIDE always present (unless rock rather low in Ca + Mg in which case hornblende forms; diopside may also be absent in rocks extremely high in Ca + Mg, e.g. in forsterite - bearing marbles)
...ent / less or absent	accessory ORTHOCLASE present unless rock is extremely low in K_2O
T / ITE	GROSSULARITE mostly present | in hottest katazone no grossularite
...ent / nearly always present	
... FELDSPAR often present	
...E commonly present	
O / CLASE	BASIC PLAGIOCLASE (mostly bytownite) always present except in some rocks extremely high in Ca + Mg
...nt / nearly always present	
...OZOIS- / ZOISITE commonly present	WOLLASTONITE frequently present
...E / EPIDOTE often present	
...ent / CLINOZ. less common	
... always present	CALCITE present if Ca+Mg > Si, or if reaction incomplete
...ays present	QUARTZ present if Ca+Mg < Si, or if reaction incomplete
SCAPOLITE often present	HORNBLENDE present if rock rather low in Ca + Mg ("conditionally stable")
...ocks (dolomitic shales)	in high Ca + Mg - excess rocks (impure marbles)
...LENDE frequently present	FORSTERITE, HUMITE, SPINEL, TREMOL., PHLOG.
...to exclusion of mica	
FELDSPAR PORPHYRO- / BLAST SCHIST / AUGENGNEISS / BANDED GNEISS	GRANITIC GNEISS or GNEISSOSE GRANITE

..., apatite, tourmaline, orthite, etc. are omitted

gneisses by synkinematic metamorphic transformation and replacement of sediments, concurrent with certain chemical changes which chiefly consisted of the addition of alkali. This problem cannot be approached without a previous discussion of the progressive metamorphism the non-granitized sediments have undergone. Moreover, granitization should not be visualized as an isolated phenomenon, but in its integral relationship to higher-grade regional metamorphism.

Most of the sediments in the Nanga Parbat area are argillites belonging to the Salkhala group, and the sedimentary base of the granitic gneisses almost entirely consists of these rocks. Therefore, the series of progressively metamorphosed argillites is the one with which we are chiefly concerned. It shows, in a continuous succession observed both across and along the strike, *a number of "mineral facies"* (cf. P. Eskola, 1920) which can be readily assigned to the three classical *"depth" zones.* The writer does not know another area in which the zones of regional metamorphism are displayed with equal completeness and continuity and in equally magnificent exposures. The *factor controlling the spatial distribution of mineral facies and zones* in this area can be shown to be exclusively *temperature.* Confining pressure was approximately the same in places greatly differing in their degree of metamorphism. For the metamorphic zones succeed each other horizontally, with approximately vertical boundaries; in fact, most of the katazonal rocks forming the interior of the Nanga Parbat massif are exposed at higher elevations than most of the meso- and epizonal rocks of the marginal areas. Stress action was also of the same intensity in the different metamorphic zones as is shown by the structures of the rocks although, as is to be expected, crystallization has healed cataclastic mineral structures more completely in the hotter inner portions of the massif than in the marginal areas.

The series of metamorphosed argillites begins with black carbonaceous slates and ends with high-grade sillimanite-bearing rocks. The upper epizone is represented by carbonaceous *phyllitic slates* composed of sericite and quartz, often with additional chlorite. They pass into less finely textured normal *epizonal phyllites* (cf. pl. 1, fig. 1) composed of the same minerals, often with the addition of albite; the albite

tends to form small porphyroblasts which frequently show helicitic structures and sometimes bear signs of having been rolled. In the albite-free low-grade rocks, some sodium is probably camouflaged in sericite (admixture of paragonite). In the *lower part of the epizone*, small porphyroblasts of supposedly manganese-rich garnet appear, still in association with chlorite.[9] At the *base of the epizone*, the phyllites become more highly crystalline and approach micaschists. The essential minerals are muscovite and quartz. Chlorite still survives as a contemporaneous (not retrogressive) constituent and assigns these rocks to the epizone although members of the biotite series occur in some of them as a predominantly fine-grained constituent resembling sericite in texture. The albite may pass into oligoclase-albite. Garnet is common. In the various epizonal Salkhala argillites described, calcite, epidote and clinozoisite are only locally found, since these argillites are usually poor in calcium. On the other hand, these minerals are common in calcareous slates and phyllites of the Cretaceous-Eocene volcanic formation. Otherwise, those rocks show the same mineral assemblages as the Salkhala argillites.

As the mesozone is entered, contemporaneous chlorite disappears. The *upper part of the mesozone* is characterized by schists composed of well-formed biotite and muscovite, and of quartz (cf. pl. 1, fig. 2). Occasionally muscovite becomes subordinate to biotite or even disappears, obviously owing to a relatively high content in iron and magnesium, and to the lack of a pronounced aluminum excess. Almandite is a common accessory and tends to form porphyroblasts; "snow-ball garnets" are common and indicate, the same as other structural features, that crystalloblastic growth and mechanical deformation were simultaneous. There is often some additional oligoclase (usually on the sodic side, owing to a low calcium content of the sediment). If the amount of oligoclase is increased, the micaschists pass into paragneisses (cf. pl. 2, fig. 1), according to the definition of the term given above. Potash-feldspar appears within the mesozone, but not at its upper boundary. None of the feldspars show porphyroblastic tendencies in these micaschists and paragneisses. The

[9] The writer cannot follow the widely used classification "chlorite zone," "biotite zone," "garnet zone," since in the rocks studied by him the ranges of stability of these minerals overlap.

amount of feldspar rarely reaches that typical of the granitic gneisses.

The *lower mesozone* is characterized by the appearance of kyanite. This mineral does not occur in all argillites metamorphosed in this zone, but only in those which have a pronounced aluminum excess, that is, in rocks derived from rather pure clays. In these rocks, kyanite often becomes one of the main constituents (cf. pl. 2, fig. 2). The other minerals are the same as in the upper part of the mesozone. However, there seems to be a tendency to somewhat reduce the amount of mica and especially of muscovite, and accordingly to increase the amount of feldspar. The crystallization of free aluminum-silicate accounts for this, most of the aluminum excess having been fixed in muscovite before the temperature had become high enough to permit formation of kyanite.[10] Accordingly, most of the kyanite rocks are paragneisses rather than micaschists, and potash-feldspar is rarely absent. Occasionally muscovite is reduced to the role of an accessory or is even absent. However, biotite-free varieties exceptionally occur; they are rocks very low in magnesium and with a moderate iron content which was fixed in the almandite present in almost all the kyanite rocks. In exceptional cases, kyanite rocks are so low in feldspar that they must be classified as micaschists, the limited amount of potash present having all been used up by the mica, and the quantity of sodium having only been sufficient to make some accessory oligoclase. Apart from kyanite rocks, the lower mesozone contains "ordinary" micaschists and paragneisses hardly distinguishable from similar rocks of the upper mesozone (cf. pl. 2, fig. 1).

In the *katazone*, *sillimanite* takes the place of kyanite (cf. pl. 3, fig. 1). All sillimanite-rich rocks collected were found to contain a considerable amount of potash feldspar usually represented by sanidine, and generally a considerably lower amount of oligoclase. At the same time, biotite considerably exceeds muscovite, and this latter may even be absent; high temperature which led to the crystallization of almost all of the aluminum excess as sillimanite, is the obvious explanation.

[10] It is of course realized that the amount of water available is another factor controlling the relative amounts of mica and feldspar. However, this factor can here be neglected since there are indications that there was no shortage of water in any of the metamorphic zones in this area.

Like the kyanite-rich rocks, those rich in sillimanite are derived from rather pure clays. Some of the sillimanite-rich rocks are practically free of quartz. Along with the high aluminum excess, this underlines the pure nature of the original clays from which this particular variety of rock was derived.

As a whole, non-granitized metamorphosed argillites occur only in very subordinate relict bands in the katazonal interior of the Nanga Parbat massif. Most of these bands do not contain sillimanite, but are "ordinary" paragneisses. Their mineral composition does not differ from that of the "ordinary" paragneisses of the mesozone, except for frequent absence of muscovite.[11] As in the mesozone, the oligoclase is usually on the sodic side. This is due to a primary low calcium content. Where katazonal paragneisses are in contact with lime-silicate bands, their plagioclase becomes at once more basic. The low calcium content of the meso- and katazonal paragneisses is also demonstrated by the presence of rutile whereas in metamorphosed calcareous shales of both zones the titanium has combined with calcium to form sphene. Like the mesozonally metamorphosed argillites, those of the katazone lack porphyroblastic development in their feldspars. In all these rocks, the carbonaceous matter characteristic of the Salkhala slates is still widely present in the form of finely scattered graphite flakes.

The textures of the meso- and katazonally metamorphosed argillites are fine-grained. Most of these rocks are dark-colored, owing to their high content in finely divided biotite. The appearance of the metamorphosed argillites is very different from that of the granitized rocks which show light colors, coarser textures, and porphyroblastic feldspars.

Lime-silicate series. — The argillite series is perfectly matched by the series of progressively metamorphosed impure calcareous rocks occurring as subordinate thin bands in the Salkhalas and, in the eastern portion of the area, in the lower part of the Cretaceous-Eocene volcanic formation. Considering the small amount of calcareous rocks in the Salkhalas, and in view of the fact that they have usually escaped granitiza-

[11] The writer cannot agree with the opinion held by many authors that muscovite is a stranger to the katazone. He thinks that muscovite, like hornblende, is "*unconditionally stable*" in the mesozone, and is "*conditionally stable*" in the katazone (meaning, stable under certain limited chemical conditions). This subject will be treated elsewhere.

tion, their metamorphism is only briefly discussed in the present paper.[12]

Upper-epizonal calcareous slates and *epizonal calc-phyllites* contain the same minerals as the non-calcareous argillites, with additional calcite (dolomite), epidote and clinozoisite. Also here, fine-grained members of the biotite series appear before chlorite becomes unstable. Calc-micaschists with either biotite or muscovite or both, are frequently rich in epidote or zoisite; they range close to the boundary of epi- and mesozone, and seem to belong to the highest mesozone. To the same mineral facies belong *calc-micaschists with hornblende porphyroblasts*; part of them are developed as *garbenschiefer*. These rocks are derived from dolomitic shales whereas the epidote-zoisite-rich micaschists represent calcareous shales. Varieties rich in magnesium and very poor in potash become para-amphibolites free of mica. In some of these rocks, the green hornblende is still associated with contemporaneous chlorite which assigns this variety to the lowest epizone.

The majority of the rocks just described belong to the Cretaceous-Eocene volcanic formation although bands of calcareous phyllites and micaschists are also found in the Salkhalas, often together with marble layers. In the *upper mesozone*, such bands are characteristically developed as *calcareous zoisite-bearing biotite-muscovite-quartz-schists*; they usually contain oligoclase and often almandite. Within the *lower mesozone*—corresponding to the kyanite-schist facies in the pure argillites—scapolite appears in the impure calcareous bands intercalated in the Salkhalas. The rocks are *scapolite-paragneisses* and, more rarely, scapolite-micaschists. Their minerals are: scapolite, zoisite, biotite, muscovite (often absent), oligoclase, potash-feldspar, calcite, quartz, almandite, and sphene. The latter mineral is common in impure calcareous rocks from epi- to katazone. Impure epi- and mesozonal marbles contain the same minerals as the schists described.

In the katazone, the association calcite + quartz becomes unstable in the impure calcareous bands in the Salkhalas. Now true lime-silicate minerals form. However, not infrequently the reaction remains incomplete, presumably owing to the CO_2 being unable to escape (cf. V. M. Goldschmidt, 1911). In the upper katazone, basic plagioclase and diopside appear

[12] This subject will be more fully discussed in another paper.

and scapolite still survives from the lower mesozone (unless it is retrogressive). Examples are: hornblende-bytownite-granulite with quartz, calcite and garnet, forming a band in biotite-paragneiss; diopside-scapolite-marble and granulite, with basic plagioclase and some potash-feldspar, occurring in a similar way.

With further increase in temperature, the *katazonal lime-silicate-granulite bands* reach their typical development and wollastonite is added as a characteristic constituent, representing the same mineral facies in which sillimanite forms in pure argillites. The wollastonite is associated with bytownite or bytownite-anorthite. This paragenesis is remarkable since, according to V. M. Goldschmidt's classical memoir on the contact metamorphism in the Oslo region (1911), this association could not be stable and the two above minerals ought to react to form grossularite. However, in the Nanga Parbat rocks wollastonite and bytownite are intergrown as contemporaneous and obviously stable partners. Since pressure would favor the reaction, and since the Nanga Parbat rocks have undoubtedly formed under considerably higher pressure than the Oslo and similar contact rocks, it must be concluded that in the interior of the Nanga Parbat massif *temperature was considerably higher than in the "inner contact zone,"* and was thus able to overcome the not inconsiderable pressure and prevent the grossularite reaction. This is confirmed by the presence, in part of the Nanga Parbat lime-silicate-granulites, of later grossularite replacing the paragenesis wollastonite + bytownite. The grossularite reaction began to take place when temperature, though still katazonal, had become somewhat lower.

The magnesium of the lime-silicate-granulite bands is fixed in diopside. Where such bands are in contact with ordinary paragneiss, a boundary zone rich in common green hornblende usually intervenes, demonstrating that hornblende, containing less Ca as well as Ca + Mg than diopside, is "conditionally stable" in the katazone (cf. above, foot-note no. 11). As the rock grades into purer argillite, biotite takes the place of hornblende. In the lime-silicate bands, biotite never occurs, and the potash is fixed in some accessory orthoclase. The lime-silicate-granulites contain either excess quartz or excess calcite, but sometimes both are present (incomplete reaction,

cf. above). Bands of *lime-silicate-marble* often contain forsterite, instead of diopside (most economic use of limited silica available). If aluminum was available and there was a shortage of silica, spinel has formed in marble bands. Also, phlogopite is found in marbles of this zone.

The gradation, biotite-paragneiss, hornblende-granulite, lime-silicate-granulite, lime-silicate-marble, pure marble, is often observed within a few centimeters and even less, and close and regular interbanding is common. It is thus indicated that the primary differences in chemical composition which can be traced back to but slightly metamorphosed Salkhala sediments, have been perfectly preserved even under katazonal conditions, and that *practically no chemical migrations have taken place within the calcareous rocks.* This fact is remarkable, in view of the considerable chemical changes which have transformed most of the metamorphosed argillites into granitic gneisses (cf. below). The preservation of pure marble layers, not only next to paragneiss, but even within (migmatitic) granitic gneiss, illustrates the lack of chemical migrations in the calcareous bands.

The progressive metamorphism of argillaceous and impure calcareous rocks at Nanga Parbat is summarized in table 1. The *progressive metamorphism of basic rocks* matches that of the sediments. Basic rocks occur in the Cretaceous-Eocene volcanic formation, in the norite masses, and, as subordinate intercalations, in the Salkhalas and the gneisses formed from them. There is no need to describe in the present paper the various mineral facies displayed by the basic rocks. It may only be mentioned that in the sillimanite-gneiss and lime-silicate-granulite facies, there is no sign of an "eclogite facies" in the basic rocks, but they are developed as katazonal amphibolites; obviously, the eclogite facies is not a normal katazonal facies, but one restricted to extreme pressure, e. g. great depth.

(III) SYNKINEMATIC GRANITIZATION IN NANGA PARBAT MASSIF

(*a*) *Outer zones*

After metamorphism of the Salkhala argillites had reached a certain stage within the lower mesozone, their granitization began. Its intensity steadily increased toward the katazonal interior of the gneiss massif.

Both on the eastern and western side of the Nanga Parbat massif, granitization set in within the kyanite zone. First smaller and then large porphyroblasts of potash-feldspar began to form in the fine-grained schists (cf. pl. 3, fig. 2). As the number of porphyroblasts increases, the *feldspar-porphyroblast-schists* grade into *coarse-grained augengneisses* of granitic appearance (cf. pl. 4, fig. 1) and, if porphyroblasts are aligned in stringers and coalesce into bands, they grade into *banded gneisses* (cf. pl. 3, fig. 2; pl. 4, figs. 2-4). Both kinds of metamorphic migmatites are usually rich in almandite. There are all kinds of gradations, intermediate types, interlayering and interfingering of these variously granitized rocks, as well as of non-granitized metamorphic argillites.

(1) *Mesozonal migmatitic augengneisses.* — Both on the eastern and western side of the Nanga Parbat massif, there is a *continuous belt* thousands of feet wide in which massive coarse-grained augengneisses greatly predominate. These belts occupy the outermost parts of the massif of granitized Salkhalas. This augengneiss belt is absent only in the southeastern part of the area shown in figure 1.

Although having an igneous appearance (pl. 4, fig. 1), these massive augengneisses reveal their metasomatic origin in the field, by the presence of all gradations between augengneisses and argillite-derived schists and in the interlayering of both kinds of rock (cf. Misch, 1935). Microscopic examination confirms the metasomatic-migmatitic nature of the augengneisses. Their *groundmass* is, in its mineral composition and texture, identical with the biotite-schists and paragneisses described above which are altered Salkhala argillites. In this matrix, the large *potash-feldspar porphyroblasts*, usually microperthite, grew, replaced and penetrated it, and enclosed relics of its mineral grains (pl. 5, fig. 1; pl. 6, fig. 2; pl. 7, fig. 1). The irregular and crenulated borders of the porphyroblasts finally approach crystal faces, as the porphyroblasts tend to become idioblastic (pl. 6, fig. 2). Karlsbad twinning is common. The amount of porphyroblasts become very large, they exceed half of the volume of many rocks, and the original schist becomes restricted to relict stringers and winding bands between the large feldspars. Of particular interest is the presence of kyanite in the argillite-derived matrix of many of the augengneisses

(pl. 5, fig. 1; pl. 7, fig. 1). Projections of growing feldspar porphyroblasts can be seen invading kyanite-rich paragneiss fabric, and absorbing it (pl. 7, fig. 1). Pure aluminum-silicate would hardly crystallize in a rock so rich in potash that more than half of it consists of potash-feldspar (not to mention the additional potash fixed in mica). In other words, the kyanite chemically represents an *unstable relic*. One chemical equilibrium was succeeded by a new and different one, and the difference was chiefly due to an introduction of potash.

The texture of the rock indicates that the potash-feldspar porphyroblasts have not been introduced as a whole, but are the product of a *reaction between the argillite-derived schist and newly introduced material* rich in potash. The introduced material was in a condition which enabled it to infiltrate and permeate the country rock, and to react with it. This means that the introduced material consisted of *hot solutions*.[13] The formation of the potash-feldspar porphyroblasts cannot be attributed to metamorphic differentiation within the country rock, both because the amount of potash in the augengneisses is higher than in the argillite-derived schists in the non-granitized areas, and because the matrix of the augengneisses possesses the original mineral composition of the argillite-derived schists and cannot therefore be a differentiate.

The *amount of material added* was not large (no exact figures can be given unless chemical analyses are made). A normal argillite is not too far from a granite in its chemical composition, and an addition of a moderate amount of alkali and usually some silica will transform it into a rock of granitic composition; whether silica is needed depends on the amount of quartzose silt admixed with clay minerals in the original argillites. In the mineral composition of these augengneisses,

[13] Several authors have recently emphasized the importance of solid diffusion in chemical migrations during metamorphism. Though it has been demonstrated that such a process is possible, the writer's observations lead him to believe that solid diffusion will, if at all, usually take place only on a very minor scale (as between adjacent minerals—albitization of basic plagioclase where invaded by newly formed hornblende, observed by the writer in metanorites), but can hardly be held responsible for major chemical transfers during regional metamorphism and granitization; the action of solutions is often shown by textural features connected with the intergranular film, and that water was present is indicated by the occurrence of hydrous silicates, as well as by general considerations (water content of sediments, especially of argillites, phyllites, micaschists, etc.).

there is no indication of an introduction of sodium at this stage. However, there is a possibility that some sodium was introduced already at an earlier stage of metamorphism, namely in those mesozonal paragneisses, which contain a considerable amount of oligoclase. Since these rocks are fine-grained and lack porphyroblastic feldspars, there is no textural indication of an introduction of sodium, and it could be ascertained only by chemical analyses. Even if there has been some addition of sodium, these fine-grained and even-textured paragneisses could not possibly be described as "granitized."

Unless an equal amount of material has been removed in solution from the country rock—of which there is no indication in the rock successions in question—the *volume* must have been somewhat increased by granitization. Such a small increase in volume would not be easy to prove even in a rock granitized under static conditions, and is much more difficult to ascertain in these rocks which have been metamorphosed and granitized during mechanical deformation. The winding shape of bands of paragneiss material between the feldspar porphyroblasts might be considered as an indication of increase of volume in the areas replaced by porphyroblasts. That the augengneisses are *synkinematic* is shown by the elongation and alignment the porphyroblasts have acquired during their crystalloblastic growth, and by cataclastic deformation the porphyroblasts have suffered at some places where movement outlasted growth, whereas in other places all cataclastic structures have been healed by crystallization.

(2) *Mesozonal migmatic banded gneisses.*—Banded gneisses occur as intercalations within the belts of massive augengneiss, but have their main development on the inner side of these belts. Here they form *zones several miles wide* on both sides of the Nanga Parbat massif. In the southeast of the area where the augengneiss belt disappears, the zone of banded gneisses grades into the area of non-granitized argillite-derived schists. The banded gneisses consist of alternating layers of dark-colored biotite-schist and paragneiss, and of white or light-colored gneiss of the appearance of an acidic orthogneiss (pl. 3, fig. 2; pl. 4, figs. 2-4).

The *biotite-schist and paragneiss bands* are, in composition and texture, identical with the rocks occurring in the non-

granitized areas of metamorphic Salkhala argillites. Kyanite frequently occurs in these bands (pl. 5, fig. 2; pl. 6, fig. 1). The *light-colored bands* (pl. 7, fig. 2; pl. 8) are for the most part less fine-grained, and are leucocratic and rich in feldspar, especially potash-feldspar (frequently microcline). Muscovite is generally present in addition to potash-feldspar, oligoclase (usually sodic), and quartz. Biotite occurs in much smaller amounts than in the dark-colored bands, and is often altogether absent. Almandite and tourmaline are common accessories both in the light- and dark-colored bands. Tourmaline is also widely developed in the non-granitized areas of metamorphic argillites, and therefore can hardly be considered as a "pneumatolytic" mineral, but may have been derived from the boron originally contained in the marine argillites.

Apart from white bands with little or no biotite, there are also *bands intermediate in composition* and color between the dark-colored schists and the white bands. All gradations and all kinds of interbanding occur (pl. 4, figs. 2 and 3). The *thickness of individual bands* of dark-colored schist and light-colored gneiss is widely variable, both ranging from a few millimeters to many feet. The *proportion* of both kinds of rock in the banded gneisses is likewise extremely variable. The series begins with metamorphic argillites containing only a few thin white bands usually accompanied by individual feldspar porphyroblasts (pl. 3, fig. 2), and it ends with white acidic gneisses in which only subordinate thin layers of schist have been preserved (pl. 8). Banded gneisses showing different proportions are intercalated in numerous repetitions.

There occur *all passages between banded gneisses, augengneisses, and feldspar-porphyroblast-schists* (pl. 3, fig. 2; pl. 4, fig. 4; pl. 5, fig. 2; pl. 6, fig. 1), and subordinate layers of typical augengneisses may be intercalated in the main zone of banded gneisses. Across the strike, light-colored bands frequently grade into schists with potash-feldspar porphyroblasts (pl. 4, fig. 4). Along the strike, light-colored bands frequently die out within the schist, or are continued by layers rich in potash-feldspar porphyroblasts (pl. 4, figs. 2 and 4). Looking at such layers from the other end, their porphyroblasts may be said to coalesce finally into white bands. Also within the light-colored bands, the feldspar frequently shows a *porphyroblastic development*, often discernible in the field

PLATE 1

Fig. 1. Epizonally metamorphic argillite: phyllite. Sericite and quartz. Plain polarized light. North of Kalapani, southeastern part of Nanga Parbat area (specimen no. 16.8.24).

Fig. 2. Mesozonally metamorphic argillite: micaschist. Muscovite, biotite, quartz, almandite. Plain polarized light. Rampur ridge, eastern part of Nanga Parbat area (no. 19.6.25).

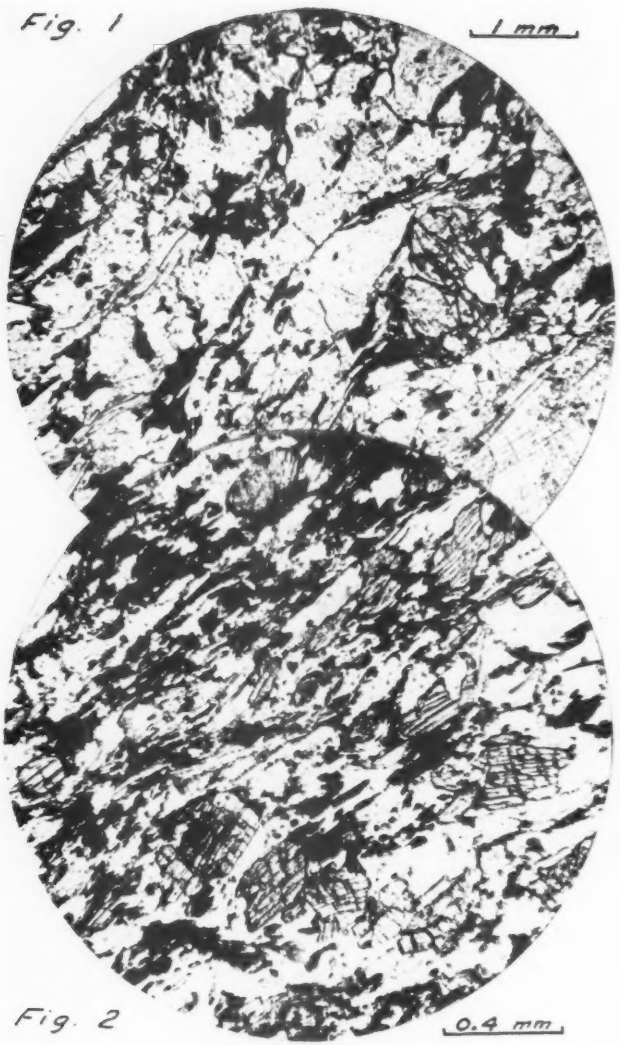


PLATE 2

Fig. 1. Mesozonally metamorphic argillite: biotite-paragneiss. Biotite, muscovite (less), oligoclase, quartz, almandite. Plain polarized light. Bunar Nullah, northwestern part of Nanga Parbat area (no. 24.5).

Fig. 2. "Low"-mesozonally metamorphosed argillite: kyanite-biotite-paragneiss. Kyanite, biotite, oligoclase, potash-feldspar, quartz, almandite, traces of muscovite. Plain polarized light. Ridge south of Rama Valley, eastern part of Nanga Parbat area (no. 12.6.2).

PLATE 3

Fig. 1. Katazonally metamorphic argillite: sillimanite-biotite-paragneiss. Quartz-less variety, portion rich in sillimanite. Sillimanite, biotite, orthoclase (sanidine), oligoclase (less). Plain polarized light. North face of Nanga Parbat above Ganalo Glacier, central part of Nanga Parbat gneiss massif (no. 24.7.8).

Fig. 2. Beginning of granitization in metamorphic argillite: mesozonal banded feldspar-porphyroblast-gneiss. Rock surface, nat. size. South of Mangdoyan north of Astor, eastern part of Nanga Parbat gneiss massif (no. 13.5.8a).

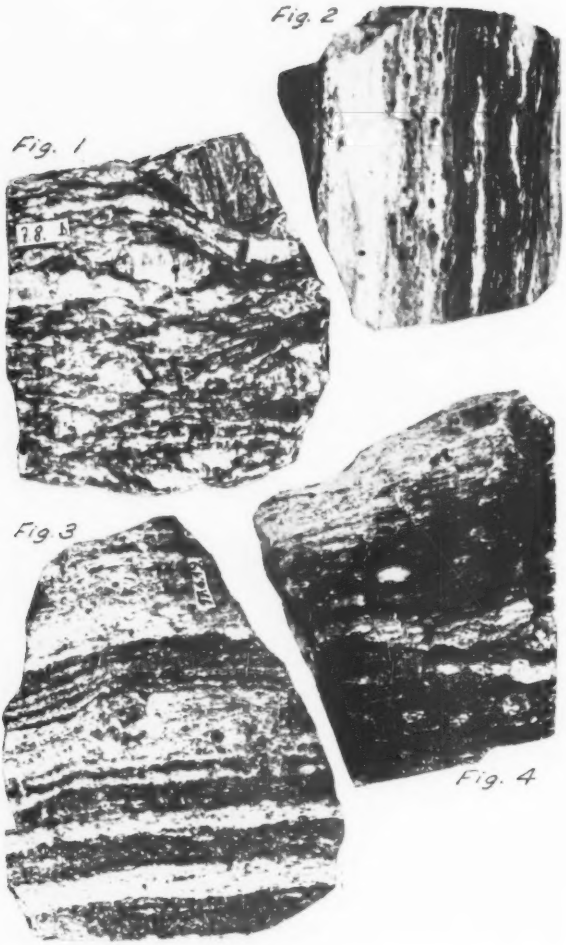


PLATE 4

Granitization in mesozonal metamorphic argillites.

Fig. 1. Augengneiss. Potash-feldspar porphyroblasts replacing fine-grained dark colored kyanite-biotite-paragneiss. Abt. $\frac{1}{2}$ nat. size. South of Astor, eastern part of Nanga Parbat gneiss massif (no. 7.8b).

Fig. 2. Banded gneiss (tourmaline-bearing). Completely granitized white quartz-feldspar bands, non-replaced dark-colored biotite-paragneiss bands, and intermediate bands. Thinning of white replacement bands. Small potash-feldspar porphyroblasts. About $\frac{1}{2}$ nat. size. North of Astor, eastern part of Nanga Parbat gneiss massif (no. 9.6.5).

Fig. 3. Banded gneiss. Gradations between light-colored replacement layers and dark-colored biotite-paragneiss bands. In replacement layers some porphyroblastic texture megascopically visible. About $\frac{1}{2}$ nat. size. Chugam area south of Astor, eastern part of Nanga Parbat massif (no. 17.6.19).

Fig. 4. Banded gneiss showing passage to augengneiss. Potash-feldspar porphyroblasts in part coalesce into replacement bands. About $\frac{1}{2}$ nat. size. Gishat Nullah, southeastern part of Nanga Parbat area (no. 17.8.8a).

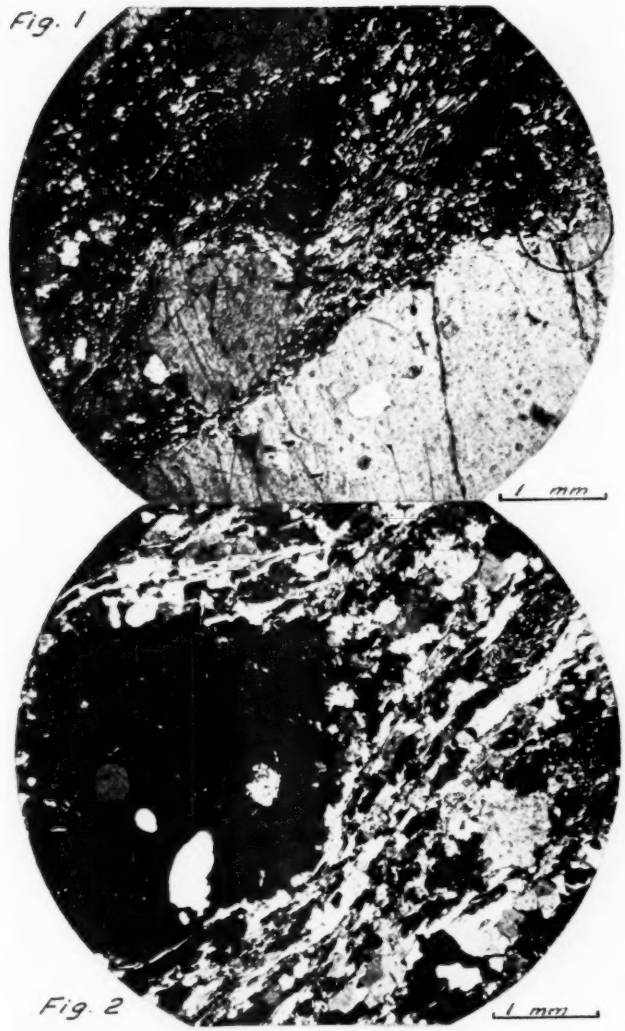


PLATE 5

Fig. 1. Mesozonal augengneiss. Microperthite porphyroblasts metasomatically replacing kyanite-biotite-paragneiss. Crossed nicols. Circle on right: area shown in plate 7, fig. 1. South of Astor, eastern part of Nanga Parbat gneiss massif (no. 10.6. augengn.-valley).

Fig. 2. Potash-feldspar porphyroblast replacing kyanite- and biotite-bearing groundmass in incompletely replaced light-colored banded gneiss of mesozone. Crossed nicols. Ridge north of Rama Valley north of Astor, eastern part of Nanga Parbat gneiss massif (no. 11.6.9).

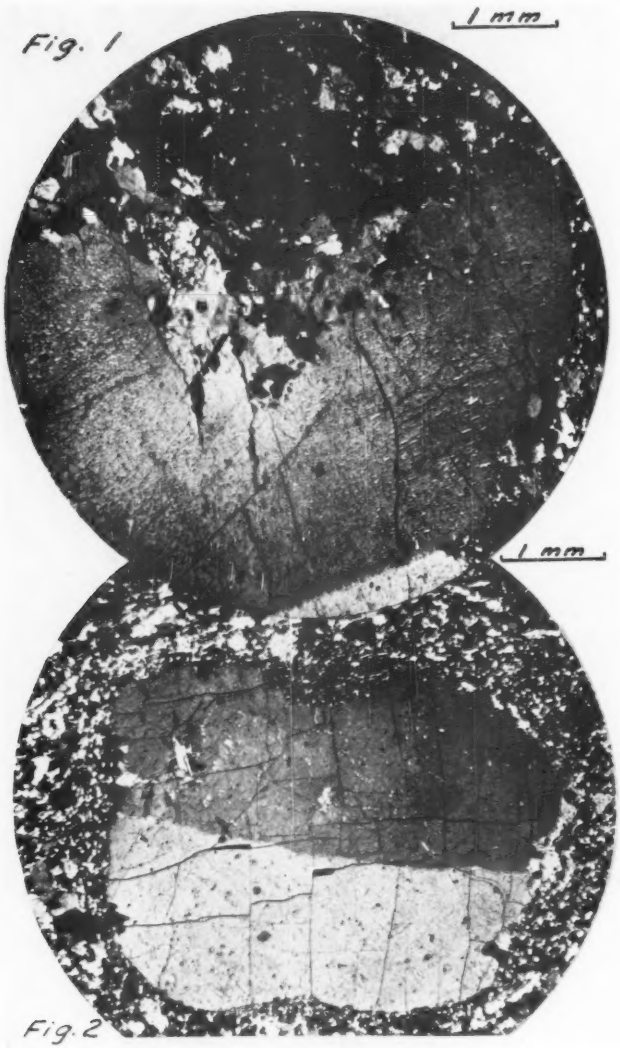


PLATE 6

Fig. 1. Part of large microperthite porphyroblast replacing fine-grained kyanite-biotite-paragneiss in porphyroblastic and partly lenticular banded gneiss of mesozone. Crossed nicols. South of Astor, eastern part of Nanga Parbat gneiss massif (no. 7.8.g).

Fig. 2. Microperthite porphyroblast replacing very fine-grained biotite-paragneiss. Mesozonal augengneiss comparatively poor in augen (passage to feldspar-porphyroblast-schist). Idioblastic tendency of porphyroblast. Crossed nicols. South of Astor, eastern part of Nanga Parbat gneiss massif (no. 7.8.d).

PLATE 7

Fig. 1. Border of microperthite porphyroblast replacing kyanite- and biotite-rich paragneiss matrix in mesozonal augengneiss. Detailed view of area marked by circle in pl. 5, fig. 1. Crossed nicols. South of Astor (no. 10.6. augengn.-valley).

Fig. 2. Completely granitized white band in mesozonal banded gneiss. Porphyroblastic microcline, quartz, oligoclase, very little muscovite. Granoblastic. Crossed nicols. Dichil Finger, eastern border zone of Nanga Parbat gneiss massif north of Astor (no. 3.7.37).

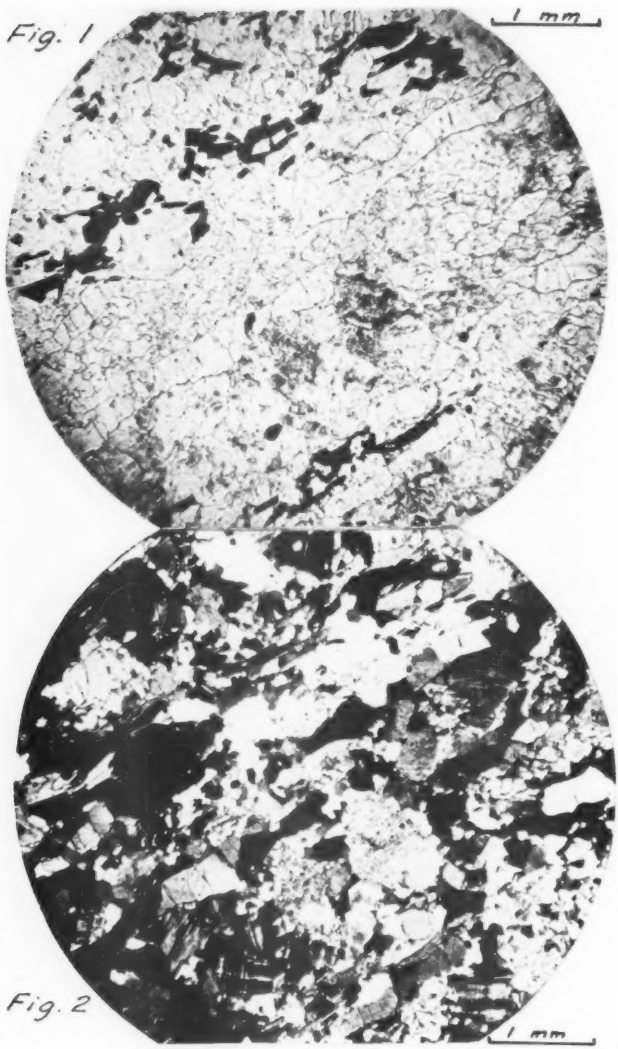


PLATE 8

Figs. 1 and 2. Mesozonal banded gneiss. Portion showing high degree of replacement. Relict bands of biotite-paragneiss being invaded and absorbed. Olive-green biotite, microcline, quartz, oligoclase, in biotitic bands accessory epidote with orthite. Figs 1 and 2 show the same area, in plain polarized light and under crossed nicols. East ridge of P. 16,000′ west of Rattu valley, eastern part of Nanga Parbat gneiss massif (no. 21.6.8).

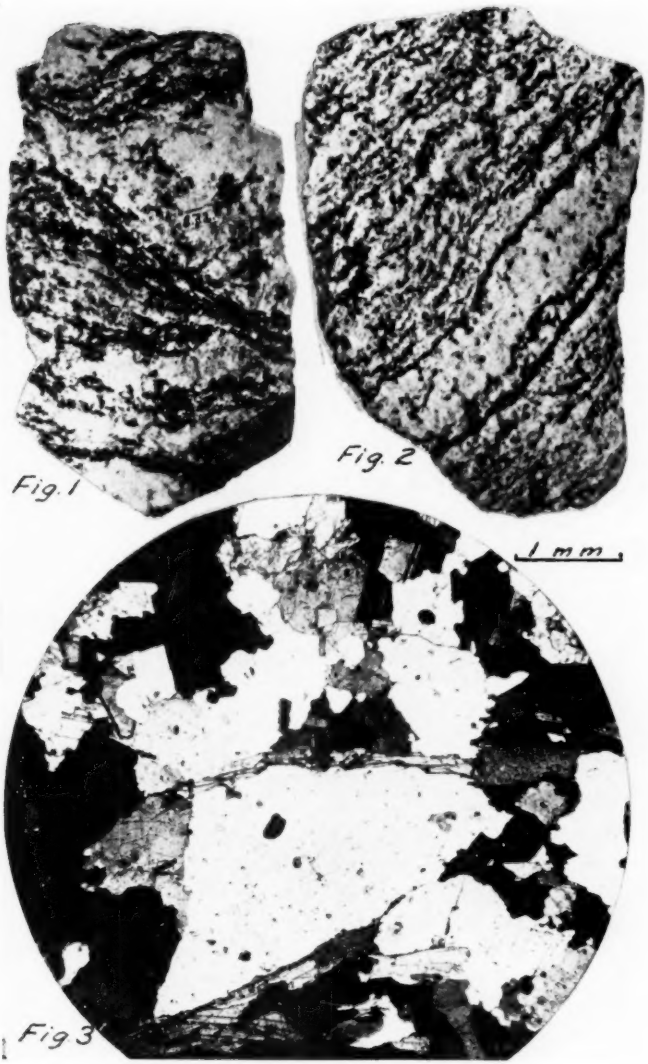


PLATE 9

Katazonal granitic gneiss forming interior of Nanga Parbat gneiss massif.

Fig. 1. Very coarse-grained biotite-flasergneiss. Uneven texture. Porphyroblastic growth of feldspar. About ½ nat. size. Buldar ridge northeast of Panga Parbat (no. 8.7.2).

Fig. 2. Banded biotite-flasergneiss. About ⅔ nat. size. North face of Nanga Parbat (no. 18.7.2c).

Fig. 3. Coarse-grained biotite-flasergneiss. Orthoclase, oligoclase, quartz, biotite, some almandite. Final stage of granitization. Granoblastic texture. Crossed nicols. North face of Nanga Parbat (no. 18.7.1).

(pl. 4, fig. 3), and more generally visible under the microscope (pl. 5, fig. 2; pl. 6, fig. 1; pl. 7, fig. 2). At the borders of light-colored bands, the feldspars can often be seen invading the schist (in a manner comparable to that observed in the augengneisses), and thereby to "dilute" the schist so that its biotite becomes more thinly spread (pl. 8). Occasionally, more coarsely recrystallized biotite becomes concentrated in extremely thin layers (almost films) along sharp borders of white bands.

The similarity in composition of the dark-colored bands and the metamorphic Salkhala argillites of non-granitized areas, and the great variability in the thickness and proportion of dark- and light-colored bands compared with the slightly variable composition of the non-granitized schists, prove that *the banding cannot be due to metamorphic differentiation* of the argillite-derived country rock, but has been brought about by an introduction of material. The transformation of the metamorphic argillites into banded gneisses has been accompanied by a *definite change in the chemical composition of the rock*, and at the end of this series of transformations occur those banded gneisses in which unaltered schist material has become very subordinate. Metamorphic differentiation has taken place only very locally and on a very minor scale, producing those thin biotite-enriched bands which have been occasionally segregated at the borders of white bands.[14] Judging from the mineral composition of the rocks, the introduced material in the banded gneisses is alkali (predominantly potash), and usually also silica. The possibility that some sodium had been introduced already at an earlier stage of metamorphism, has been discussed above (cf. chapter 1).

Which was the *mechanism of introduction* of this material? Banded gneisses of the type described are commonly considered as the result of "*lit-par-lit injection.*" There are considerable

[14] In the Nanga Parbat area, metamorphic differentiation appears to be responsible for the formation of certain banded amphibolites derived from some of the norites and related rocks. In this case, lighter and darker bands are composed of exactly the same mineral varieties, and only their relative amounts differ in different bands. There is no indication that the bulk composition of the rock has been changed. The individual bands are uniformly thin, and the proportion of lighter and darker bands does not vary greatly. I consider these features as characteristic of banding originated by metamorphic differentiation, in contrast to banding due to introduction of foreign material.

divergencies in the use of the term "injection." In this country, it is usually applied in its restricted and proper sense, e. g. for an intrusion of magma or magmatic solutions or similar mobile materials into some space created by a breaking-open or splitting-up of the solid country rock, emplacement being a mechanical process. In Europe, the term "injection" has been widely used in a looser and more comprehensive sense, comprising processes of infiltration, soaking and replacement, as well as true injection, but excluding intrusion of larger continuous bodies of magma. The term has, for instance been used in this sense by V. M. Goldschmidt in his classical study of the "injection metamorphism in the Stavanger region" in Norway (Goldschmidt, 1921), in which part of the "injection gneisses" described are shown to owe their origin to infiltration and replacement by solutions (assumed to be derived from a magmatic source). More recently, C. E. Wegmann has, in an important contribution to the analysis of migmatites (Wegmann, 1935), suggested the restriction of the term "injection" to its proper meaning. The definition given by him agrees with the usage customary in this country; with the modification that intrusion of larger continuous bodies of magma is excluded from the term "injection." In a preliminary report on the Nanga Parbat area written immediately after my return from the field (Misch, 1935), I have used the term "injection" in the broad sense which used to be customary in Europe, but I would now apply it only in its restricted sense. In that paper, I applied the term "injection gneisses" to various types of gneisses of metamorphic-migmatic origin, including the porphyroblast-gneisses and the massive augengneisses, although their replacement origin was pointed out in that paper. The banded gneisses I then considered as the result of true lit-par-lit injection by acidic-granitic material derived from a magmatic source. Subsequent petrographic study of the rock suites from Nanga Parbat has, however, convinced me that most, and probably all, of the material composing the light-colored layers in the banded gneisses owes its origin to *infiltration and replacement of the country rock, rather than to true injection.*

The following features which have been described above, are in favor of such an interpretation by "*lit-par-lit replacement*": the presence of numerous bands intermediate in com-

position between the original schists and the white gneiss bands, and the occurrence of gradational contacts—features which suggest different degrees of replacement of the original schist; the intimate association of banded gneiss with, and its grading into, undoubted replacement rocks, e.g. augengneiss and feldspar-porphyroblast-schist; and the porphyroblastic texture widely developed in light-colored gneiss bands, which links such bands still more closely to layers of porphyroblasts that often prolong the light-colored bands. There is also a general argument against an origin of the banded gneisses by true lit-par-lit injection. It is difficult to visualize that schists while being subjected to considerable pressure and to action of intense orogenic stress, could have been split open by magma or magmatic solutions throughout areas measuring tens of miles along the strike and a number of miles across it. On the other hand, infiltration along layers of active shearing could have taken place under such mechanical conditions. This argument seems to be even more convincing where banded gneisses do not dip steeply, but have gentle dips, as in certain parts of the Nanga Parbat area,[15] and as in some other regions I have since had an opportunity to study. If in such cases banding were the result of true injection, each injected layer of liquid would have had to uplift and carry the tremendous load of thick overlying rock masses.

The suggestion that lit-par-lit replacement has played the main part in the formation of the banded gneisses, does not preclude the possibility that there may have been some subordinate contribution of true injection material which became activated as a final result of the migmatic metamorphism. However, no structural features indicative of mass movement of granitic material were observed. On the contrary, the tectonic structure of the banded gneisses—as well as of all other migmatitic gneisses in the Nanga Parbat area—is of a type characteristic of deformation of solid rocks during orogenic metamorphism, and is identical with the type of structure seen in the non-migmatitic schist areas. This general structural aspect of the problem will be discussed below.

Lit-par-lit replacement, as against true lit-par-lit injection, implies an emphasis on *chemical processes*, as against a purely mechanical process. The introduced material infiltrated and

[15] Parts of the southeast and northeast; cf. figs. 17 and 21 in (Misch, 1935).

permeated the country rock, and reacted with it. To make banded gneiss, the *loci of infiltration and reaction* must have been thin layers in some way distinguished from neighboring layers. One might think of primary differences in the mineral composition of the schists, certain layers being more pervious or having a greater chemical affinity to the granitizing solutions.[16] However, the areas of non-migmatitic metamorphosed argillites do not show primary differences of this kind, and in the southeastern part of the Nanga Parbat region these areas of metamorphosed argillites grade into banded gneisses and augengneisses along the strike (cf. fig. 1). The explanation must then lie in some other condition that existed during the time of migmatitic metamorphism. It appears that this was the condition of active shearing movements along the foliation planes of the country rock which took place during the regional metamorphism and the migmatitization connected with it. The introduced material, consisting of solutions,[17] infiltrated the country rock *along layers distinguished as particularly active shear zones* during the differential movements which were taking place throughout the country rock. The loci of layers of feldspar porphyroblasts were determined in the same manner as those of the white bands eventually formed.

There arises the question why the result of migmatitic infiltration should have been predominantly massive augengneiss in certain marginal belts, and chiefly banded gneiss in neighboring zones. These two rock varieties cannot be considered merely as less and more advanced stages in the process of migmatitic transformation; for the banded gneisses contain layers of non-migmatitic schist whereas such are, apart from subordinate intercalations, absent in the augengneiss belts. Incipient banded gneisses are represented by schists with layers of porphyroblasts, but not by uniformly feldspathized rocks. The difference between the belts of augengneiss and the zones of banded gneiss are probably due to a more highly volatile character of the infiltrating solutions in the marginal belts predominantly transformed into augengneiss, the solutions thus being able to permeate the country rock in a more uni-

[16] The latter argument applies wherein the Nanga Parbat massif calcareous or basic intercalations have, in contrast to the enclosing metamorphic argillites, escaped granitization.

[17] As to the solutional character of the introduced material, I refer to the discussion with respect to augengneiss.

form manner. Such a disposition seems plausible at a time when the "migmatite front" (to use a term coined by Wegmann, 1935) had advanced to its maximum extent and remained more or less stationary for a period.

It has been stated above that the banded gneisses were formed when orogenic stress was active, which means under conditions of differential movement within the rock masses concerned. Foliated structure within the light-colored as well as the schist bands, general sharp parallelism of banding and foliation and of other elements such as stringers of porphyroblasts, and presence of some cataclastic mineral structures not healed, prove the *synkinematic* (synorogenic) character of the banded gneisses.

(*b*) *Migmatitic gneissose granites in the interior of the Nanga Parbat massif*

Though the augengneisses and banded gneisses in the outer zones of the Nanga Parbat gneiss massif have been shown to owe their origin to metasomatic granitization, it is still feasible that these zones, though having a width of miles and occupying an area of more than 100 square miles, could be nothing but extensive contact zones granitized along the margins of a synorogenic intrusive body occupying many hundreds of square miles in the interior of the massif. We cannot therefore speak of "metasomatic granitization of batholithic dimensions," unless it also can be demonstrated that the granitic gneiss in the interior of the massif owes its origin to granitization, and not to igneous intrusion.

The rocks forming the interior of the gneiss massif have been outlined in chapter (II), and the non-granitized metamorphic sediments occurring here briefly described in chapter (III). Whereas in the banded gneisses and augengneisses of the marginal areas large quantities of argillite-derived fine-grained paragneiss and biotite-schist material are preserved and are easily distinguished from the metasomatically replaced portions of the rocks, this distinction becomes less well marked in the interior of the massif, and typical fine-grained paragneiss material is preserved only in quantitatively very subordinate intercalations. Otherwise, the argillite-derived rocks gradually become more coarsely recrystallized and feldspar-enriched, as the interior of the massif is approached. The result is mainly flasergneisses of mod-

erately fine to medium grain and of medium shades. Their sedimentary derivation is only shown by their grading into typical fine-grained paragneisses (cf. Misch, 1935). Toward the interior of the massif, these rocks pass into the *coarse-grained biotite-flasergneiss of granitic appearance and composition* (*or gneissose granite*) which here predominates (cf. pl. 9, figs. 1 and 2).

This gneiss is composed of orthoclase, oligoclase (varying from a little over 10 to somewhat over 20 percent anorthite), quartz, biotite, and frequently also muscovite as an equal and contemporaneous constituent. Almandite is a common accessory. Although the rock is rather uniform regionally, it is anything but homogeneous in detail, and its *texture is usually very uneven*. The gneiss is generally coarse-grained though the grain size ranges from more medium-grained to extremely coarse-grained varieties, often within very short distances. Some of the gneisses are even-granular, but an uneven grain, due to the development of large feldspars (especially orthoclase), is more common and occurs always in extremely coarse-grained varieties. The biotite content of the gneiss varies greatly, even within short distances. In single rock specimens the biotite is usually unevenly distributed, being concentrated in bands, stringers and patches (pl. 9, figs. 1 and 2). The almandite may also be enriched in certain bands. The gneisses possess crystallization-foliation (free of cataclastic mineral deformation), but the degree of their schistosity is variable and never becomes extreme. Most of the rocks show flaser structure which may pass into augen structure (cf. pl. 9, fig. 1). More typical augengneisses are subordinate. They differ in their texture from those of the marginal parts of the Nanga Parbat massif, their groundmass being more coarse-grained and their orthoclase porphyroblasts usually being less well-formed crystals. In some augengneisses, the biotite-rich groundmass exceeds the porphyroblasts. Banded structures are quite common in the gneisses, and are often found in light-colored varieties (pl. 9, fig. 2). Acidic bands are occasionally bordered by thin layers enriched in biotite (local metamorphic differentiation). The banded and augen structures described recall those typical of the marginal zones of the massif, with coarse-grained crystallization of the entire rock and more general feldspathization superimposed. Banded and augengneisses of the type of the marginal zones, characterized by

the preservation of fine-grained and non-granitized schist material, are very rare in the interior of the massif; they occur only occasionally in association with the subordinate intercalations of fine-grained paragneiss found in the granitic gneiss. The microscopic texture of the granitic gneiss is crystalloblastic and usually shows the same uneven character which is megascopically visible; it can be described as granoblastic, partly with porphyroblastic tendencies. Many gneisses show a tendency to approach an almost igneous-appearing microscopic texture which many petrographers would classify as "pan-allotriomorphic granular," but the crystalloblastic character and the uneven nature of the fabric remain preserved. An advanced case is shown in plate 9, figure 3.

The character of the gneiss is strongly suggestive of a metamorphic - migmatitic rather than an intrusive origin. Among the features described, the following are especially indicative of such an origin: variability of composition, grain size and other textural features; uneven nature of texture; frequent presence of porphyroblastic features; crystalloblastic texture (though this latter does not constitute conclusive evidence since it could be superimposed). Of particular interest are the augen and banded structures described above; they suggest that these rocks had originally gone through the stage of mesozonal incomplete granitization which is now displayed in the marginal zones of the massif, and had subsequently, with further spreading of granitization, become more coarsely recrystallized and more thoroughly feldspathized, so that the paragneiss material still preserved in the mesozonal banded and augengneisses lost its distinctive character, although the banded and augen structure is preserved as a relic. In general, the biotite-rich stringers and patches in the unevenly textured granitic gneisses are rather suggestive of recrystallized relics of biotite-paragneiss. Exceptionally, sillimanite was found to occur in granitic gneiss, indicating participation of argillite-derived material.

Apart from the character of the gneiss itself, there are *other and even stronger arguments in favor of an origin by metasomatic granitization.* The first is the presence of *intercalations of undoubted metamorphic sediments* and some basic rocks which, though quantitatively subordinate, occur throughout the wide area of granitic gneiss. *These intercalations are structurally conformable with the granitic gneiss.* Most of

them are thin, and many of them can be followed for long distances. It is difficult to visualize that mechanical intrusion of magma could have spared such septa and, even if it did, could have left them structurally intact without thoroughly disturbing and disrupting them. If, on the other hand, these intercalations are interpreted as relict bands which have escaped metasomatic granitization, their presence and their mode of occurrence are readily understood.

Many of these intercalations consist of katazonal fine-grained paragneiss whose contacts with the granitic gneiss are usually gradational. Some layers within such paragneiss intercalations may be partially granitized, having been transformed into feldspar-porphyroblast-schist and porphyroblastic augen-gneiss. Often paragneiss and granitic gneiss alternate in parallel layers. All these features indicate granitic replacement rather than intrusion.

Of particular interest are the *thin marble layers* with their associated lime-silicate-granulite bands, which form regular conformable intercalations in the gneiss often traceable for long distances. None of the marble layers observed exceed a thickness of 100 feet and most of them are much thinner. No intrusion could have left these bands structurally intact; moreover, it is difficult to comprehend how the marbles could have escaped fusion if temperature had been high enough to keep silicate compounds in a molten state at the existing pressure for a considerable time in large magmatic masses surrounding the marbles. There is more marble relative to metamorphic argillite in the interior of the gneiss massif than in the area of non-granitized Salkhala phyllite and of schist and paragneiss derived from it. The absolute proportion of marble, on the other hand, is nearly of the same order in the interior of the massif as in the non-granitized areas. This proves that most of the argillites which greatly predominate in the original sedimentary section, became transformed into granitic gneiss whereas most, if not all, of the marbles escaped granitization—a truly metasomatic process, and one of differential character. In this connection, it should be recalled that the non-granitized Salkhala argillites grade into the mesozonal migmatitic gneisses and then into the central granitic gneisses not only across but also along the strike. Generalizing, it may be said that the marbles extend from an area of phyllites through one of schists into one of mesozonally migmatitized schists and con-

tinue right into the granitic gneisses. The relict nature of the thin marble bands is particularly striking where they occur as conformable intercalations in coarse-grained granitic gneiss without being accompanied by any paragneiss, all of the argillite originally enclosing the marble having been transformed into granitic gneiss. Several cases of this kind have been observed (cf. fig. 2), although the majority of the marbles are still associated with some paragneiss which varies in thickness from a few inches to several hundred feet.

The reason why calcareous bands have escaped granitization is obvious. Whereas it does not take much introduced material to transform an argillite into a rock of granitic composition, the substance of a marble would have to be almost entirely replaced to achieve the same result. Apart from the different chemical affinity of marble and argillite to granite, their relative permeability to alkali-rich solutions is probably an additional factor. Whereas slate is undoubtedly less pervious to aqueous solutions than limestone, their metamorphic equivalents must be supposed to act differently under conditions of granitization. Now metamorphic argillite would be more pervious, consisting of a granular fabric of silicate minerals (and being, in case of synkinematic granitization, in a state of active differential movement); and, having an excess of aluminum, metamorphic argillite would have a strong tendency to precipitate alkali from permeating solutions rich in alkali. On the other hand, under the existing high pressure preventing escape of its CO_2, marble would be less pervious to those solutions; at the same time, it would have no tendency of precipitating alkali from solutions rich in it. Also, under pressure, marble might plastically flow and hence be less permeable than schist traversed by active shearing planes.

Basic rocks—chiefly katazonal amphibolites—intercalated in the interior of the gneiss massif, and originally forming bands in the Salkhala argillites, have likewise essentially escaped granitization. Generally, granitization of basic rocks is not uncommon, and has been described by a number of authors. In areas where I have observed this process, there is usually the following series of transformation: original basic rock (amphibolite, basic hornfels, etc.; sometimes with lime-silicate rock at the beginning of the series, other times with metabasalt, greenstone, metanorite, metagabbro, etc.); gabbroid rock; dioritic rock; quartz-dioritic and, not always,

granodioritic rock (for instance: Indus Valley north of Nanga Parbat; Fish Lake, Palmer Lake and other places in Okanogan country, Eastern Washington; Northern Sierra Nevada; all these examples are postkinematic). In this series, the alkali added apart from silica etc. is sodium—in contrast to the predominant potash metasomatism in the Nanga Parbat gneiss massif—and as far as potash has at all been introduced, it has come in only near the end of the transformation. There can be no doubt that there are two chief chemical types of granitization, i.e. *predominant potash metasomatism*, and *predominant sodium metasomatism*. Both are apt to act on metamorphosed argillites, but the latter seems to be more easily able to attack basic rocks apart from argillites. The only effect of granitization of the potash metasomatism type on basic intercalations which I have observed in the Nanga Parbat gneiss massif, consists of biotitization of hornblende which has often marginally transformed hornblende-rich intercalations into almost pure schistose biotite rock.

Summing up from a structural angle, the relationships described can be said to be characterized by *structural continuity*. This applies if we proceed from the non-granitized zones through the incompletely granitized to the more fully granitized areas, and is also valid in a stratigraphic sense, insofar as individual stratigraphic units run through these three zones. Structural continuity also exists within the granitized areas between granitized rocks and non-granitized relics (regularity of mutual intercalation, conformable and gradational relations, harmonic and equal participation in the same individual folds).

The structural features listed confirm the replacement origin of the granitic gneiss, and illustrate the absence of intrusive structures. The same is demonstrated by the *type of tectonic deformation*. It is identical in the zone of phyllites, the area of non-granitized schists, the mesozonal migmatitic belts, and the katazonal region of granitic gneiss with its relict intercalations. The structure in all these zones is characterized by well-defined trends which remain constant for long distances and are clearly related to the regional structure of the northwest Himalayas, and by a regular character of the major folds. There is no sign of that structural turmoil and turbulent folding and flowage which one would expect if magma had broken into the sediments under high pressure and intense

stress, and which is known to occur even in solid rocks deformed under conditions of extreme plasticity. There are, of course, as in all areas of crystalline schists, numerous intensely compressed minor folds of "mobile" character throughout the zones described, but they are, in the interior of the gneiss massif as well as in the phyllite area, all regularly aligned parallel to the prevailing local trend. Most of these minor folds are isoclinal, and their narrow arcs can often be seen to become obliterated by the foliation which is parallel to their limbs.

The type of structure described is that *typical of solid rocks deformed under conditions of regional metamorphism.* The fact that this type of deformation characterizes the granitic gneiss area as well as the phyllite zone, is the last argument I wish to offer in favor of a replacement origin of the granitic gneiss or gneissose granite, and therewith of the whole of the vast gneiss massif of Nanga Parbat.

A short discussion on the *chemical composition of the "granitizing solutions"* is necessary. The *predominant solvent* was undoubtedly water (presence of hydrous silicates in all metamorphic zones; water content of the original rocks). As to the substances carried, *material present in a solution* and *material precipitated from a solution* must be distinguished. Only the latter is *directly recorded* in the rocks. If, for instance, an aluminum-excess rock has precipitated potash from a solution and augengneiss has thus been formed, the permeating solution may have contained other substances not recorded by precipitation. However, certain *indirect conclusions* are possible. The solution from which potash was precipitated must have had an excess of alkali over aluminum, if it did at all contain more than traces of aluminum. Moreover, the solution must have contained more potash than sodium. Otherwise, the porphyroblasts of microperthite in the augengneiss would be accompanied by such of sodic plagioclase. On the other hand, the solution may have contained a large or a smaller amount of silica. Silica would not be necessarily precipitated by a rock already possessing an excess of free silica. When kyanite and mica of the original schist were replaced by feldspar porphyroblasts, some silica was consumed. Some silica was supplied by quartz of the original schist, which together with mica and kyanite was replaced by feldspar. Where the amount of quartz was not sufficient, addi-

tional silica was precipitated from the solution. Whether a small amount of silica has been added in the augengneisses, could be ascertained only by a series of chemical analyses. In the white replacement layers in the banded gneisses silica has undoubtedly been introduced in addition to alkali. For these layers are rich both in feldspar and quartz, and poor in mica, and thus more silicic than the original mica-rich schist.

Another indirect conclusion as to the composition of the granitizing solutions can be drawn from the plagioclase in the granitized rocks. The mesozonal augengneisses and banded gneisses usually contain sodic oligoclase, like the non-granitized metamorphic argillites. If the granitizing solutions had contained any considerable amounts of calcium, the plagioclase in the mesozonal granitized rocks would have become a calcic oligoclase, and possibly epidote minerals would have formed. Even the katazonal migmatitic granitic gneisses usually contain a sodic oligoclase. If the granitizing solutions had contained even a moderate amount of calcium, the plagioclase would of necessity have become more calcic under katazonal temperatures. If the solutions had had any considerable content in iron (and magnesium), the amount of biotite could not show a marked decrease in the granitized rocks compared to non-granitized schists, and the intercalated marbles would probably have been affected.

This leads to another criterion. Where granitizing solutions have acted *differentially* on rocks of different composition, further indirect conclusions as to the chemical nature of the solutions are possible. If in the same section rock A has been granitized and rock B has escaped granitization, the permeating solutions were obviously rich in those substances required to transform rock A into one of granitic composition, but deficient in those elements primarily needed to convert rock B into one of granitic composition. Solutions which were "granitizing" with regard to rock A, do not deserve this name with regard to rock B. Differential granitization therefore makes it possible to eliminate a number of substances as essential constituents of the solution.

An example is granitization of basic rocks, such as amphibolites, in comparison to that of metamorphic argillites. It has been shown above that basic rocks are mainly apt to be granitized if the permeating solutions have a high sodium content—apart from the large amount of silica which is of

course required. It seems that predominant potash metasomatism does not generally affect basic rocks very much (apart from biotitization), although potash is frequently introduced at a late stage of the transformation of basic rocks, whereas the earlier stages are usually characterized by definite sodium predominance.

(IV) RELATIONS BETWEEN PROGRESSIVE REGIONAL METAMORPHISM AND SYNKINEMATIC GRANITIZATION

From a general point of view, one of the most striking features displayed in the Nanga Parbat area is the *intimate relationship and the simultaneous increase of progressive regional metamorphism and of synkinematic granitization.* Throughout the area, granitization does not begin before regional metamorphism in the argillites has reached the stage of the low-mesozonal kyanite-schist facies, and it invariably does start in this zone. It here remains rather incomplete—considerable quantities of non-replaced argillite material being preserved—and its type is that represented by banded and augengneisses. As regional metamorphism approaches the stage of the katazonal sillimanite-paragneiss and lime-silicate-granulite facies, granitization becomes more complete—argillite relics being subordinate—and develops the type represented by granitic gneiss or gneissose granite.

The succession of zones and mineral facies of regional metamorphism can in this area be shown to be exclusively *a function of increasing temperature* (cf. chapter II a). Since the degrees of regional metamorphism and of synkinematic granitization are definitely linked, this latter must likewise increase with temperature, and the same must apply to the introduction of material which is implied in metasomatic granitization. In other words, *rise of temperature and increasing infiltration by granitizing solutions go parallel.*

The rise of temperature during metamorphism and granitization in this, and many other areas, is *unrelated to geologic depth*, i. e. is not a normal geothermal rise.[18] There must, therefore, have been a *differential supply of heat* resulting in an upward curving of geisotherms—at very steep angles in this area in which the metamorphic isograds are approxi-

[18] Normal geothermal rise of temperature can, of course, also reach temperatures sufficiently high for high-grade regional metamorphism, but *only at great depths* of the crust. Here, migmatitization may take place without introduction of additional heat.

mately vertical. There are no known processes which could have created a sufficiently large amount of heat *in situ*. Heat must therefore have been *introduced* from below. Was this done by conduction or convection? The former seems unlikely in itself, considering the high angle of upward curvature of geisotherms; the systematic relation between rise of geisotherms and introduction of granitizing solutions confirms that *convection was the means of introduction of heat*. In other words, *the granitizing solutions carried both chemical substances and heat*. It is in these terms that we have to conceive the "rising of the migmatite front," to use again the phrase coined by Wegmann (1935). If no regional introduction of material were involved in granitization, it could be argued that progressive metamorphism, being itself controlled by rise of temperature, is the cause and granitization merely a consequence. However, material has undoubtedly been added "from somewhere below"; to demonstrate this fact, the chemical changes involved in the making of the migmatitic gneisses have been stressed, and it has been pointed out why these gneisses cannot have formed by processes of metamorphic differentiation of the country rock, i. e. by local transfer as against regional introduction of material.

If the analysis given above is correct, *high-grade regional metamorphism is not only necessarily linked with synkinematic granitization, but depends on it for a supply of additional heat, with the exception of great depths of the crust*. In the Nanga Parbat area, regional metamorphism without granitization did not go beyond the phyllitic (and perhaps uppermost mesozonal) stage. All zones of higher-grade metamorphism are in their spatial distribution so clearly related to the central rise of the migmatite front, that a genetic control is obvious. This applies not only to the high-grade metamorphic sediments preserved within the granitized area, but also to the zone of higher-grade rocks outside the margin of the granitized area. The rocks involved are the typically mesozonal (higher mesozone and beginning of lower mesozone) mica-schists and paragneisses which form a belt between the low-grade rocks and the mesozonal augengneisses and banded gneisses. To this belt, additional heat must have been supplied from the neighboring granitized area, and the means may have been conduction, since the rocks in question are outside the granitized

area and heat could not therefore have been conveyed to them by the granitizing potash-rich solutions directly—unless this belt was, outside the area of actual granitization, reached by aqueous solutions carrying some sodium (cf. above). To generalize, *the front of synorogenic granitization is in time and space preceded by a zone of regional metamorphism of a higher grade than that achieved by orogenic metamorphism not linked with granitization.* It is obvious that *the further increase* (from low-mesozonal to katazonal) *of the metamorphic grade in the sediments preserved within the granitized area of Nanga Parbat, illustrates the control by successively increased introduction of heat, conveyed by successively more active and hotter granitizing solutions.* In the granitized rocks themselves, this progression is shown quantitatively by increasing completeness of granitization, and qualitatively by the passage to a more truly granitic type of gneiss in the interior of the massif.

The *Nanga Parbat example* of progressive metamorphism and synkinematic granitization is *typical* and, it may be said, *normal*; its only distinguishing features are unusually clear, deep and continuous exposures. The rocks and processes described are repeated all over the world. Provided the principles suggested are correct at Nanga Parbat, they will also apply elsewhere, subject, of course, to certain modifications due to varying physical and chemical conditions (cf. below). However, one qualification must be made. Synorogenic granitization and metamorphism at Nanga Parbat took place in a *geosynclinal environment.* Therefore, areas which can be properly compared are other geosynclines. Metamorphism and granitization are characteristic of orogenic transformation in the interior portions of first-order geosynclines, and they here often rise to *comparatively high levels. At extreme depths* of the crust, processes will be modified. Such cases are believed to be displayed in some Archeozoic regions, and perhaps in the roots of some exceptionally deeply eroded later geosynclinal chains. Numerous areas of Archeozoic crystalline schists do not, however, differ from such of later ages, and can frequently be shown to represent old geosynclines exposed at levels comparable to those visible in later geosynclines.

The statement made above concerning the Nanga Parbat

area, that *orogenic metamorphism without granitization* is apt to achieve only low-grade mineral facies ("pure dynamometamorphism") is borne out by numerous migmatite-free regions of phyllites and basal-epizonal and upper-mesozonal micaschists, and of other low-grade schists (greenschists, etc.). Among regions I have studied, a good example is the geosyncline of West Yunnan at the southwestern border of China (cf. Misch, 1945 a and b). Here Paleozoic and Mesozoic sediments and volcanics have been dynamically metamorphosed during Mesozoic orogeny into epizonal and occasionally uppermost-mesozonal crystalline schists within an area of many thousands of square miles, without having achieved a higher grade of metamorphism anywhere. No synkinematic migmatites of Mesozoic age have been observed anywhere in this region.

The *beginning of granitization* of the potash-metasomatism type was at Nanga Parbat found to be within the lower mesozone, and granitization was shown to be preceded in space and in time by mesozonal regional metamorphism resulting from an introduction of additional heat from the area of granitization (possibly connected with some finely divided sodium-precipitation from solutions). These relations seem to be general; for in all areas of synorogenic migmatites known to me, metamorphism has achieved a mesozonal mineral facies before granitization has set in. However, there appears to be some variation as to the part of the mesozone, and therewith as to the temperature, at which granitization begins. There seem to be cases where migmatic processes start already higher within the mesozone. A good example seems to be that of the Stavanger region in Norway described by V. M. Goldschmidt (1921). Here synorogenic granitization is predominantly of the sodium-metasomatism type. It is entirely of this type in the zone of beginning replacement (albite- and oligoclase-porphyroblast-schists); this zone is upper-mesozonal, judging from the minerals described in the schists. It appears, from the rock succession recorded by Goldschmidt, that here additional potash-metasomatism came in only at a more advanced stage of progressive granitization.[19] There seem to be other

[19] The more typically granitic rocks in this region are described as intrusive, but the continuous nature of the series is so pronounced, and the general character of the rock succession so strikingly resembles that described in the present paper, that an identical genetic interpretation suggests itself.

cases which also suggest that synorogenic granitization of the sodium-metasomatism type is apt to begin at a temperature lower than that at which potash-metasomatism was found to commence at Nanga Parbat.

Finally, I wish to emphasize *the general validity of the statement that high-grade regional metamorphism and synorogenic granitization are invariably and necessarily linked.* They are expressions of one and the same process. There is not any region known to me either from study of the literature or from personal field work, where katazonal or lower-mesozonal crystalline schists occur without metamorphic migmatites or granitic rocks.

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(*To be Continued*)

REGULARITY OF OLD FAITHFUL GEYSER, YELLOWSTONE NATIONAL PARK, WYOMING[1]

PHILIP F. FIX[2]

ABSTRACT. It is common belief that old Faithful Geyser, the most celebrated geyser in the world, (1) has become irregular and even erratic, (2) does not erupt in winter, (3) erupts to heights much less than formerly, (4) discharges much less water than formerly, and (5) is the only regular large geyser.

This erroneous belief is based upon (1) inaccurate observation, (2) incorrect interpretation and correlation of data, (3) failure to use standard nomenclature, (4) second-hand opinions, and (5) unbridled romanticism.

All available data of reasonable reliability prove that (1) these popular legends are false, (2) the record for 1870-1947 shows a degree of relative regularity in behavior that is amazing for a geyser of such great size and complexity of function, (3) the amazing stability of the average interval between the initiations of eruptions, observed as 65.1 minutes in 1878 and as 65.12 minutes for the entire period 1870-1947, is unequalled, (4) it cannot be shown that since discovery in 1870 the behavior of the geyser has changed fundamentally in any way, although its idiosyncrasies have become better known, and (5) the same fallacious statements were being made in 1879. Old Faithful Geyser richly deserves its world-wide fame.

Introduction

OLD Faithful Geyser, the most celebrated geyser in the world, is in the Upper Geyser Basin in the valley of the Firehole River in the southwest-central part of Yellowstone National Park, Wyoming. Although discovery and naming of the geyser are credited to the Washburn-Langford-Doane expedition of 1870 because of its monumental work in making known the magnificent features of Yellowstone, the geyser was well known to James Bridger and other far-ranging "mountain men" for many years before 1870, and to the aborigines for untold years or centuries before that time. The origin of the geyser was discussed by Bauer and Marler.[3]

The writer became deeply interested in the geyser in 1932, and in addition to observations made as time permitted through 1939, he studied it daily from early June to mid-September

[1] Published with permission of the Director, U. S. National Park Service.

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[3] Bauer, C. M., and Marler, George. Old Faithful, an example of geyser development in Yellowstone Park. Assoc. Pacific Coast Geographers Year Book, 5, 1939, pp. 45-48. Northwest Science, 13, 1939, pp. 50-55.

1937 while stationed at Old Faithful. This paper embodying results of the study before and during the time the writer was an employee of the U. S. National Park Service was prepared for publication in 1938, but the report of an extraordinarily long interval between two eruptions in that summer led him to postpone publication until additional perspective might be obtained from additional records. Observations for 1938, and by the U. S. National Park Service for 1939-1947, have been added, and all available data of acceptable accuracy therefore are correlated in this paper.

The writer is grateful to the following friends in the U. S. National Park Service for courtesies that facilitated his study: Dr. C. Max Bauer, Chief Geologist of the U. S. National Park Service; Chief Park Naturalist David DeL. Condon and Park Naturalist W. Verde Watson of Yellowstone National Park; and Carl Berryman, George Marler, Herbert T. Lystrup, W. Leon Evans, W. E. Kearns, Jennings King, Lloyd Sweetman, John Thompson, Julius Roller, and H. R. Woodward. He is grateful to J. E. Haynes, authorized Park Photographer, for the superb photographs illustrating this paper; for the record of his observations in 1910; and for many courtesies during the study.

No record of the behavior of Old Faithful Geyser would be complete without appropriate tribute to the painstaking work of A. C. Peale, whose records for parts of 1872 and 1878 and summary of the observations by other persons from 1870 to 1879[4] provide most of the available information concerning the early history of the geyser. The detail and the excellent form in which his records were kept have never been excelled. The writer found Peale's record and nomenclature so good that it was continued as part of the foundation for his own paper on geyser nomenclature.[5] Many garbled interpretations of geyser behavior have resulted from lack of appreciation and respect for the excellent prior nomenclature established by Peale, and to the lack of forethought and care.

The accurate determination of time seems to have received little attention from observers, who, by failing to standardize

[4] Peale, A. C. Thermal Springs (of Yellowstone National Park): Old Faithful Geyser. U. S. Geol. and Geog. Survey of the Territories (Hayden), 12th Ann. Rept., pt. 2, 1878 (1883), pp. 220-229.

[5] Fix, Philip F. Nomenclature of geyser eruptions. Jour. Geology, 47, 1939, pp. 99-104.

their chronometric instruments, have left us only the unfortunate record of geyser behavior as interpreted by the particular timepiece of unknown accuracy. Although facilities for accurate control of the writer's watch were not available in the summer of 1937, and it was not reset during the summer, the watch was slightly less than 3 minutes ahead at the end of the summer.

The Old Faithful Geyser of Folklore

It is commonly supposed that "in the good old days," Old Faithful Geyser erupted with the precision of clockwork. Many persons claim to have witnessed such behavior, but they have no *records* to show. Many of the old frontier men who were guides in the early years of the Park fostered such a belief as an integral part of the folklore of the Wild West then in vogue, and many visitors were so overwhelmed by the wonder of what they saw in Yellowstone that they failed completely to distinguish between fact and fiction when it was before their very eyes.

The writer has frequently heard it said that Old Faithful Geyser does not erupt at night, nor in the winter. One visitor said that on a previous visit several years before, she had seen Old Faithful erupting from another mound of sinter in the vicinity. Another said that she had seen the geyser erupt black smoke at the time of her previous visit 6 years before. The writer learned that what she actually saw was the water column of the geyser silhouetted against the early morning sun. The height of the geyser eruptions and the volume of water discharged likewise have been the subjects of unbridled imagination.

There is a fatal lack of conformity in the tales of what the geyser is supposed to have done. Some say that the eruptions occurred precisely every 60 minutes, but others claim with

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PLATE 1

Fig. 1. This world-famous photograph shows an eruption of Old Faithful Geyser on a windless day with the water curtain falling around the foot of the geyser column (Haynes Inc. photo No. 13040).

Fig. 2. Eruption of Old Faithful Geyser on an average day silhouetted against the light, with a light wind carrying the curtain of water to fall beyond the geyser column (Haynes Inc. photo No. 10161).

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equal vehemence that the eruptions occurred precisely every 61, 63, 65, 66, or 67 minutes. Baseless rumors still are being manufactured. The writer heard on a radio program in 1938 that "Old Faithful Geyser erupts every 62 seconds." Furthermore, such precision is utterly impossible for a geyser of the magnitude and complexity of Old Faithful, whose eruptions are the result of the coordinated teamwork of a very large number of factors that differ greatly in kind and degree.

The fatal lack of conformity in the popular legends proves that they are unworthy of serious consideration. Examination of the written record of observations in 1878 alone extinguishes all such claims.[6]

The Old Faithful Geyser of Fact

THE RECORD OF FREQUENCY

The feature of Old Faithful Geyser that has most intrigued the imagination of visitors in Yellowstone is the interval, which is that period of time between the initiations of two successive eruptions. In Old Faithful and other geysers of the same general type, a preliminary phase marked by the discharge of preliminary spurts of water precedes the initiation of the eruption phase proper.[7] The average interval for 1870-1947 is of greatest interest, but the length of each individual interval also is of considerable interest in geyser studies. The calculation of intervals in table 1 includes only those observations known to be reasonably accurate and whose method of calculation is known.

Table 1 does not show an interval on the night of July 13, 1938, that was reported to be 115½ minutes long, and that is generally thought to be an error that resulted from overlooking an eruption during that time. Such an interval approximately twice the length of the average interval, and far longer than any other known, occurring in the darkness of night inevitably is suspect because the writer and many others know, from deliberately awaiting eruptions of the geyser in broad daylight, how many times Old Faithful succeeds in attaining full eruption unnoticed when water-spray and sound are whisked away by the wind.

[6] Peale, A. C., op. cit., pp. 224-228.
[7] Fix, Philip F., op. cit., pp. 99-102.

TABLE 1

Interval between Initiations of Eruptions, Old Faithful Geyser, 1870-1947[a].

Year	Observer	Eruptions seen	Interval		
			Minimum	Maximum	Average
1870	Doane	....	.....	.....	50m
1870	Washburn	....	.....	.....	60m
1872	Peale	17	65m 00s	70m 34s	67m 54s
1873	Comstock	11	52m 05s	77m 35s	63m 48s
1875	Dana and Grinnell	24	.....	.....	65–66m
1877	Sherman and Poe	7	62m	80m	67m
1878	Peale	97	54m 04s	78m 00s	65m 06s
1879	Mitchell	....	.....	.....	70m
1879	Seguin	....	.....	.....	45m
1883	Hallock	....	.....	.....	65m±
1928	Allen and Day	6	57m 30s	74m	.....
1928–9	Baker	38	58m 20s	75m	.....
1932	Robertson	1187	38m	81m	65.7m
1937	Fix (July 17)	2	.....	84m	.....
1937	Fix (Aug. 25–Sept. 3)	23	42m 58s	79m 47s	67m 38s
1938	Woodward[b]	771	.....	.....	66m 30s
1939	U.S.N.P.S.	500	40m	86m	66.1m
1940	U.S.N.P.S.	1075	39m	87m	67m
1941	U.S.N.P.S.	1577	38m	91m	66.3m
1942	U.S.N.P.S.	1452	41m	90m	60.05m
1943	U.S.N.P.S.	376	39m	84.5m	63.45m
1945	U.S.N.P.S.	940	39m	85m	63.77m
1946	U.S.N.P.S.	1495	34m	86m	63.2m
1947	U.S.N.P.S.	1500	39m	89m	63.3m
1870–1947		11098	34m	91m	65.12m

(11048 used to calculate average interval)

[a] Data for 1870-1879 from A. C. Peale, op. cit., p. 229. Data for 1883-1932 from E. T. Allen and Day, A. L., Hot Springs of the Yellowstone National Park, Carnegie Inst. Washington, Publ. 466, 1935, pp. 178, 184. Observations by Robertson in 1932 apparently are on basis used by the writer in this paper. Those by Allen and Day, and presumably Baker, were based on "the quiet intervals between eruptions," a method considered inaccurate by the writer because lengths of the preliminary phase and the duration vary considerably. The writer added 4½ minutes to each of their figures to convert them approximately to the basis used in this paper. Data for 1939-1947 from personal communication by David DeL. Condon.

[b] Woodward, H. R. Observations of Old Faithful Geyser. U. S. National Park Service, Yellowstone Nature Notes, vol. 15, 1938. pp. 40-41.

The observed range between the maximum and minimum intervals, and the greatest departure from the average interval, for the entire period of record are shown in table 2.

The absolute minimum, maximum, and average intervals,

and the absolute range of Old Faithful Geyser are unknown; tables 1 and 2 show only approximate values based upon a relatively small number of observations. It is surprising to find that in the 78 years of record, while the total number of eruptions observed had increased 350-fold, the observed range between maximum and minimum intervals was only doubled! It is also very remarkable that Peale observed an average interval of 65.1^{m} from 97 eruptions seen by him in 1878, and that 70 years later the average interval calculated from 11,048 eruptions is 65.12^{m}!!

TABLE 2

Observed Range between Maximum and Minimum Intervals, and Greatest Departure from the Average Interval, Old Faithful Geyser, 1870-1947

Year	No. years observed	Eruptions observed	Observed range	Greatest departure from 1870-1947 average	
				Average - minimum	Maximum - average
1870	1[a]	?	10^{m}+	15^{m} 07^{s}	?
1873	3	40±	27^{m} 35^{s}	15^{m} 07^{s}	12^{m} 28^{s}
1879	10	160	35^{m}	20^{m} 07^{s}	14^{m} 53^{s}
1937	68	1391	46^{m}	27^{m} 07^{s}	18^{m} 53^{s}
1947	78	11048	57^{m}	31^{m} 07^{s}	25^{m} 53^{s}

[a] The total number of eruptions, and the minima and maxima observed by Doane and Washburn are not known, but are greater than the difference of 10^{m} between the average interval given by each.

THE RECORD OF HEIGHT

The measurement of height of geyser eruptions is of great interest. Their height is exceedingly difficult to estimate, and very greatly misjudged. Most persons cannot estimate accurately a distance laid out on the ground under the most familiar circumstances; much less are they able to estimate a vertical distance against the intangible blue sky. Not only is the average person unable to estimate the apparent height, but he has little understanding of the conversion of apparent into true height. It is surprising how many persons completely lack the mathematical basis for understanding an instrumental determination of height by vertical angle, yet many of these persons assert their ideas of geyser height as absolute fact. These erroneous ideas have had wide circulation, whereas the scientific record has not. The legend arose that Old Faithful Geyser no longer erupts to the great height of former years.

Inasmuch as very experienced observers are unable to estimate the exact height, the only useful data are actual instrumental measurements of height in which the method used is stated clearly. Table 3 shows the acceptable available records of heights actually measured. The earlier measurements of height are a little low in comparison with the later measurements because of a change in method used. Peale measured geyser height to the top of the unbroken column of water, and apparently the other early observers did likewise. The 1910 measurements by Haynes were "the height of the water column."[8] The heights given by Allen and Day, Baker, and the writer are known to have been to the top of the highest splashes, a more accurate method.

TABLE 3

Measured Height of Eruptions, Old Faithful Geyser, 1870-1947[a].

Year	Observer	Eruptions measured	Height of eruption (feet) Mimimum	Maximum	Average
1871	Barlow	?	?	138	?
1872	Peale	5	106	130	122
1875	Dana and Grinnell	24	?	?	115
1878	Peale[b]	11	106	139	125
1910	Haynes[c]	5±	?	?	150
1928	Allen and Day[d]	6	114	150	133
1928–9	Baker[d]	38	110	160	?
1937	Fix[e]	19	111	184	146
1871–1878		40+	106	139	119
1928–1937		63[f]	110	184	143
1870–1947 summary		108+[g]	106	184	130[g]

[a] Unless otherwise specified, data in this table are from Peale, op. cit., pp. 221-229. Peale used a gradienter, and his 1872 baseline was 100 feet long.

[b] Measurement by Ladd in 1878, considered inaccurate by Peale, are not included in this table.

[c] Haynes, J. E., personal communication.

[d] Allen, E. T., and Day, A. L., op. cit., p. 185. Measured with Klausner hypsometer. Baseline 250 (?) feet long, measured with steel tape.

[e] Measured with Brunton compass and 660 foot baseline measured with steel tape. Vertical angles believed very accurate within limitations of the instrument, as result of writer's long experience.

[f] Average calculated only from the 25 measurements for which averages are stated.

[g] Average is lowered by inclusion of older measurements made to top of water column only. Measurements for 1928-1937 made to top of highest splash of water. Average is calculated only from the 70± measurements for which the averages are given.

[8] Haynes, J. E., personal communication.

Many observers seem to have overlooked the fact that the wind decapitates the water column of a geyser by whipping away the upper part as spray. The writer's observations have led him to believe that more than 50 feet may at times be subtracted from the column of Old Faithful Geyser by the wind, but neither maximum nor average values can be estimated because of inadequate data. Variations in the height of eruptions resulting from the complexity of geysers are imperfectly understood, and they cannot be evaluated without adequate records of wind influence upon the height and operation of the geyser. The observations made by the writer in 1937 include a record of the wind.

THE RECORD OF WATER DISCHARGE

Estimates of the volume of water discharged by Old Faithful Geyser have been most romantically colored. The value generally supposed in the past seems to have been about 750,000 gallons of water per eruption. This volume, necessarily being discharged in little more than 4 minutes at a diminishing rate, means a volume of about 450 second feet. This volume is 3½ times that of the Firehole River two miles downstream from Old Faithful Geyser![9] One glance at the runoff streams from the geyser eruption shows that such volume is impossible. The reliable figure of 10,000 to 12,000 gallons per eruption, a very large volume for the circumstances, was measured in 1929 by Boyer at the request of Allen and Day.[10]

Comparative Regularity of Old Faithful Geyser

The relative regularity of Old Faithful Geyser is so astonishingly great for the period 1870-1947 that the world-wide fame accorded the geyser is deserved beyond equivocation, but the common belief that Old Faithful is the *only* regular large geyser in Yellowstone is incorrect. The degree of the regularity of several other large geysers over long periods of time is not known accurately because their lesser fame has resulted in even fewer written records than for Old Faithful. The Riverside Geyser in the summer of 1937 showed a regularity greater

[9] Allen, E. T., and Day, A. L., op. cit., p. 269.
[10] Allen, E. T., and Day, A. L., op. cit., p. 184.

than that of Old Faithful.[11] Valentine Geyser [12] in the summer of 1936 erupted with a regularity comparable with that of Old Faithful, but there have been long periods of time before and since when it was irregular or inactive because of interference from its neighbors. Daisy Geyser and several others are known to have been relatively regular over long periods also.

Summary

The known acceptable records of eruptions of Old Faithful Geyser from 1870 to 1947 present the following conclusions:

1) Each observation is a special case. The *absolute* minimum, maximum, and average intervals between initiations of eruptions, and the *absolute* range between the maximum and minimum intervals are unknown. The approximations to 1947 are as follows:

Average interval	65.12^m
Minimum interval	34^m
Maximum interval	91^m
Range (Maximum — minimum)	57^m

2) The *average interval* of Old Faithful Geyser is amazingly stable. Peale found, from the 97 observations made in 1878 that constitute the only series of observations "in the good old days" comprehensive enough to afford a reliable average, that the average interval was 65.1^m. *Seventy years later*, the average interval calculated from 11,048 eruptions is 65.12^m!!

3) No change in average height of eruptions can be shown for the period of record, 1870-1947, although the records for recent years show greater heights than formerly, the opposite of the popular legends. Exact comparison of measured heights from 1870 to 1947 is impossible because of the change in method (formerly to top of the unbroken column of water, now to the top of the highest splash of water) that makes the earlier records seem a little low.

4) Old Faithful Geyser discharges about 10,000 to 12,000

[11] Fix, Philip F. The phenomenal regularity of Riverside Geyser. U. S. National Park Service, Yellowstone Nature Notes, 17, 1940, pp. 21-22.

[12] Fix, Philip F. Valentine Geyser. Idem, 14, 1937, pp. 31-37.

gallons of water per eruption, a very large volume for the circumstances.

5) It is surprising to find that in the 78 years of record, while the total number of eruptions observed increased 350 times, the observed range between maximum and minimum intervals was merely doubled.

6) The description by Peale of an eruption of Old Faithful Geyser in 1872 or 1878 might well have been written in 1947.

7) It cannot be shown that since the discovery in 1870 the behavior of Old Faithful Geyser has changed fundamentally in any way whatsoever, although its idiosyncrasies have become better known. The evidence points conclusively to the opinion that no fundamental change occurred in the geyser from 1870 to 1947, nor probably for a very long time before 1870.

8) Future research will reveal factors, trends, and cycles at present unknown because of inadequate records. Future records should include data on weather conditions and kindred factors that unquestionably play a part in eruptions of Old Faithful Geyser.

9) Old Faithful Geyser is not the only large, relatively regular geyser in the world, but to point out that there are other large but less well known geysers that have been relatively regular for short periods of time does not detract from the world-wide fame that Old Faithful Geyser so richly deserves because of its amazing degree of regularity for so many years.

10) It is significant to note, as did Peale, that as early as 1879, if not earlier, the same fallacious statements were being made about Old Faithful Geyser as are being made today. In his report for 1879, Superintendent P. W. Norris said: "Old Faithful, though still the favorite geyser, has certainly lost prestige by an occasional prolongation of its intervals between eruptions of from 5 to 15 or 20 minutes during . . . the past season."[13] It need only be pointed out that as early as 1873, six years before the Norris report and only three years after the official discovery of the geyser, a range of 27 minutes 35 seconds had been observed. It is clear that no further comment is needed concerning the amazing regularity of Old Faithful Geyser since 1870!

[13] Peale, A. C., op. cit., p. 229.

1948 Activity

The remarkable regularity of Old Faithful Geyser is still further proved by the following data for 1948 that became available after this paper had been set in type.[14]

Number of eruptions observed	1,486
Average interval	64.1^m
Minimum interval	33^m
Maximum interval	85^m

The following summary correlates these with earlier records; the insignificant changes resulting from 1,486 additional observations is noteworthy:

Average interval, 1870-1948 (based on 12,534 eruptions)	65.01^m
" " 1870-1947 (based on 11,048 eruptions)	65.12^m
Minimum interval, 1870-1948	33^m
" " 1870-1947	34^m
Maximum interval, 1870-1948	91^m
" " 1870-1947	91^m
Range (maximum — minimum), 1870-1948	58^m
" " " 1870-1947	57^m

U. S. Geological Survey
Logan, Utah

[14] Marler, George. Geyser activity in the Old Faithful area—1948. U. S. National Park Service, Yellowstone Nature Notes, 22, 1948, p. 64.

RADIOACTIVITY OF OCEAN SEDIMENTS. VI. CONCENTRATIONS OF THE RADIO-ELEMENTS IN MARINE SEDIMENTS OF THE SOUTHERN HEMISPHERE

WILLIAM D. URRY

ABSTRACT. It has been reported in previous publications of this series that the mode of variation of the radium concentration below the ocean bottom affords a method of determining time in ocean sediments. Hitherto, these researches were confined to the northern hemisphere. Similar studies in the southern hemisphere, combined with the necessary geological and biological investigations, should provide an answer to the question of the contemporaneity of glaciation in the northern and southern hemispheres. Measurements of the radium content as a function of depth in the sediment are presented here for ocean-bottom cores secured by the U.S. Navy Antarctic Expedition of 1946-47.

INTRODUCTION

MANY measurements of the radium content of deep-ocean sediments have been reported in previous publications of this series.[1-5] A number of determinations of the radium content of "grab samples" of bottom sediments were given by Piggot.[1] Among these were measurements on samples secured in the southern hemisphere, but until the present investigation was undertaken the variation of the radium content below the surface of the ocean bottom south of the equator had not been studied.

One might inquire why interest should be evinced in comparing the radioactive relations in the deposits of the southern hemisphere with those in the sediments of the northern hemisphere. Variation of the radium content has been studied in detail for a number of cores secured in the northern hemi-

[1] Piggot, C. S., 1933. The Radium Content of Ocean-Bottom Sediments. Am. Jour. Sci., 25, 229.

[2] Piggot, C. S., and Urry, Wm. D., 1939. The Radium Content of an Ocean-Bottom Core. J. Wash. Acad. Sci., 29, 405.

[3] Piggot, C. S., and Urry, Wm. D., 1941. Radioactivity of Ocean Sediments. III. Radioactive Relations in Ocean Water and Bottom Sediments. Am. Jour. Sci., 239, 81.

[4] Piggot, C. S., and Urry, Wm. D., 1942. Radioactivity of Ocean Sediments. IV. The Radium Content of Sediments of the Cayman Trough. Am. Jour. Sci., 240, 1.

[5] Urry, Wm. D., and Piggot, C. S., 1942. Radioactivity of Ocean Sediments. V. Concentrations of the Radioelements and their Significance in Red Clay. Am. Jour. Sci., 240, 93.

sphere;[6] and there is no reason to believe that the overall relations between the members of the uranium-238 series are simply dependent on the latitude and longitude. The present results fully bear out the universal nature of these radioactive relations in all marine sediments with the possible exception of inshore deposits.

Our main interest in studying many core-samples of deep-water sediments from all the oceans arises from the fact that the lack of radioactive equilibrium in the sediments, at the moment that they are deposited, provides a method of measureing time below the ocean floor for the order of the last half-million years.[7]

The concentration of uranium in the material being deposited on the ocean floor is insufficient to support its ionium content, and, therefore, the excess ionium decreases—by 50 per cent of the remainder every 82,000 years. For practical purposes, the excess ionium disappears in 400 to 500 thousand years. Thereafter, the ionium content is constant because the uranium content can be considered inexhaustible for periods of millions of years.

No reliable method is yet available for the rapid measurement of the minute amounts of ionium scattered universally throughout the Earth's crust, but fortunately the readily determined radium suffices to provide the necessary data concerning its parent ionium. With one exception[5] (which in the light of recent results may not actually be an exception), the material depositing on the ocean floor contains less radium than the ionium can support. Consequently, the following general picture for the variation of the radium content with time, and therefore, with the depth of the buried sediment, is obtained. The radium content increases below the surface rather rapidly, unless the rate of deposition is phenomenally high, until it is in equilibrium with the ionium content. Below the depth representing the time interval necessary to establish this equilibrium the radium content diminishes only so fast as the ionium content decreases. Finally, if the material at the bottom of the core be considerably older than 500,000 years, the curve of radium content against depth should exhibit a

[6] Piggot, C. S., and Urry, Wm. D., 1941. Time Relations in Ocean Sediments. Bull. Geol. Soc. Amer. **53**, 1187.

[7] Urry, Wm. D., 1942. The Radioelements in Non-equilibrium Systems. Am. Jour. Sci., **240**, 426.

flat tail and the constant radium content is then a measure of the uranium content as well as of the ionium content, all the elements in the uranium-238 series being in radioactive equilibrium. This general picture of the variation of the radium content with the depth of the sediment is found to be true, not for a few specific examples, but for all the cores of deep-ocean sediments that have been examined. Shelf deposits remain to be investigated. In a single long core of inshore sediment in the present suite, the radium content remained extremely constant from top to bottom in agreement with the results of Sanderman and Utterback[8] working with very short cores collected in the north Pacific.

All the above phenomena associated with the variation of radium content with time beneath the ocean floor can be given a complete quantitative description with the equations governing the growth and decay of the radioelements.[7] In this manner it is possible to analyze the experimental curve of radium content as a function of depth and to assign a date to the sediment at any desired depth.

In most cores of deep-sea sediments there is a wealth of evidence of the changing geological, geophysical, climatological, and oceanographical conditions of the immediate past. Studies of the lithology and biology of cores of marine sediments have produced a record of the history of these changing conditions and this record appears to be preserved in a more complete and less disturbed state in many ways, than that revealed by studies of the continents.[9] For a satisfactory solution to many of the problems of the history of the immediate past it is essential to establish a basis for correlating the evidence for specific events, such evidence often being uncovered at widely separate localities, in very different types of sediments, and in deposits accumulating at very different rates. An outstanding example of this type of problem is the question of contemporaneity of glaciation in the northern and southern hemispheres. It was with a view to providing an answer to this question in the near future that the present work was undertaken. Obviously, there is only one sound basis for

[8] Sanderman, L. A., and Utterback, C. L., 1941. Radium Content of Ocean Bottom Sediments from the Arctic Ocean, etc. Sears Foundation: J. Mar. Res. **4**, 132.

[9] Bradley, W. H., et al., 1940. Geology and Biology of North Atlantic Deep-Sea Cores. U.S.G.S. Prof. Paper 196-A.

the correlation of the evidence; it must be dated with respect to a fixed time, preferably the present.

Presentation of the time-depth analyses of the curves of radium content against depth for these cores from the southern hemisphere would serve no useful purpose here because the results of studies of the lithology, of the foraminifera, and of other properties of these sediments are not yet available. In view of the fact that the accumulation of these necessary data may require appreciable time, it seemed advisable to report here the radioactive measurements that will eventually form the basis for dating these sediments. Furthermore, a knowledge of the concentrations of the radioelements in deep-sea sediments of all the oceans is necessary to understand the relations existing between the radioelements in ocean water and in bottom sediments. These new results will form an important addition to the body of data necessary for a consideration of the principle of dynamic equilibrium in the ocean as applied to the radioelements. Attention was drawn to the complementary nature of the Ra/U ratios in water and sediment some years ago[3]; but the transiency of the radioelements complicates the problem, and little could be accomplished with the data in existence at that time. This problem will be taken up elsewhere following publication of new determinations of the radium and uranium content of river and ocean waters. Studies of this type are of interest because they promise to provide an insight into the operation of such mechanisms as co-precipitation, selective adsorption, and concentration in organic matter.

SOURCE OF THE MATERIAL

Dr. J. L. Hough, formerly of Woods Hole Oceanographic Institution,[10] accompanied the U. S. Navy Antarctic Expedition of 1946-47 on board the U.S.C.G.C. *Northwind*. He was successful in securing ten core samples of the ocean bottom varying in length from six inches to eight and one-half feet. Six of these cores, that is, all of those exceeding five feet in length, were chosen for determination of the radium content. The cores had been sealed and preserved in Lucite liners following withdrawal of the liners from the coring tube. The

[10] Present address: Geology Department, University of Illinois, Urbana, Illinois.

liners were split longitudinally and one-half was sampled at appropriate intervals of depth. Full details concerning the apparatus employed to secure these samples, and the procedures followed in treating the samples, will undoubtedly be described by a competent authority elsewhere.

An ambitious program of many studies was planned for these cores, and, therefore, a minimum amount of material could be spared for any given part of the program. There is a further limitation in the amount of material available because it is highly desirable to work with samples that represent no more than a depth of one cm. in the sediment. The quantities employed for the radioactive measurements varied from as little as 0.3 gram near the top to 2 to 3 grams near the bottom. These amounts proved to be just sufficient to measure the radium content with the required accuracy.

The titles used to describe the sediments in this paper are tentative. They are based on preliminary observations, chiefly of color and texture, and they certainly fail to describe all the less obvious changes in the type of sediment that occur within a given core, particularly in the case of those classed as glacial marine.

The locations of the sediments sampled by these cores are shown on the charts in figure 1.

EXPERIMENTAL

The apparatus and procedure for the determination of minute quantities of radium have been fully described elsewhere.[11] Most specimens of ocean sediments lose radon on standing. The determination involves the measurement of the gaseous radon in equilibrium with radium, and, therefore, in order to insure that all the radon was collected in the ionization chamber the present samples were heated as previously described[11] for 40 to 60 minutes at 550°C in a stream of nitrogen and sealed for at least 30 days prior to measurement. At the time of measurement the gas in the storage tube was collected and added to that derived from fusing the sample. The weight of sample necessary for expressing the result was determined after drying the specimens at 110°C for 24 hours and prior to heating to 550°C. Heating to the

[11] Urry, Wm. D., and Piggot, C. S., 1941. Apparatus for Determination of Small Quantities of Radium. Am. Jour. Sci., 239, 633.

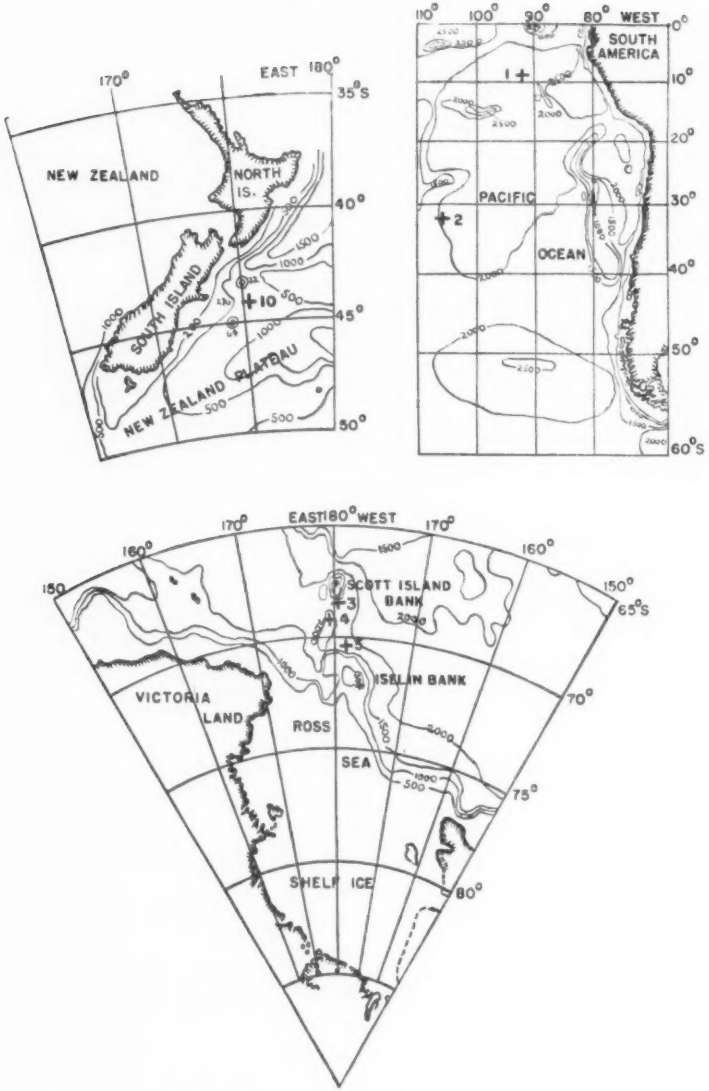

Figure 1. Partial tracings of charts of the U. S. Navy Hydrographic Office showing the location of the sites where core samples of ocean-bottom sediments were secured by the U. S. Navy Antarctic Expedition of 1946-1947. Only those cores studied for radioactivity are indicated. All depths are given in fathoms.

higher temperature avoids certain difficulties otherwise encountered in fusing the sample to liberate the radon.[11]

The root mean square probable error, insofar as the statistical fluctuations of the ion current and errors in measuring the records and setting calibration voltages are concerned, is about 1.5 per cent for most of the determinations. In a few cases where only very small samples were available the probable error was 3 per cent. Periods of observation varied from 5 to 20 hours depending upon the total amount of radium to be measured.

The absolute accuracy of the determinations is another matter. For the determination of time in ocean sediments it is only necessary to know the relative values of the radium content in any given core. However, for other purposes, such as have been indicated above, the absolute values are of interest. Radium solutions containing 10×10^{-12} g. radium, prepared by the National Bureau of Standards, were employed to calibrate the apparatus. From 1941 to 1948 repeated calibrations have always yielded the same calibration constant to within 2 per cent. This, however, affords no complete check on the agreement between various laboratories and various methods, and to provide for such checks the National Bureau of Standards prepared in 1939 a suite of rocks to serve as a means of intercalibration. With the exception of a few inhomogeneous granites, the agreement between the radium determinations at the Massachusetts Institute of Technology, the National Bureau of Standards, and the Geophysical Laboratory was satisfactory and the standard rock samples have been certified to have a given radium content within appropriate limits.[12]

In order to intercheck the apparatus used for the ocean sediments, which measures the total ion current, with another apparatus in use at the Geophysical Laboratory which determines radium by counting alpha particles,[13] one of the suite of rock samples used as standards of radioactivity has been

[11] National Research Council, 1938- . Minutes of the Committee on Standards of Radioactivity. R. D. Evans, Chairman. (The work of this committee is now the responsibility of the Sub-committee on Radiogeology of the N.R.C. Committee on Nuclear Science.)

[12] Davis, G. L., 1947. Radium Content of Ultramafic Igneous Rocks. I. Laboratory Investigation. Am. Jour. Sci., 245, 677.

measured at frequent intervals with the results shown in Table I.

TABLE I

Standard Rock Kimberlite in 10^{-12} g. radium per g.
Measurements of the Radium Content of the

NBS value		0.59 ± .04	some sealed, some not sealed
Geophysical Laboratory			
ion current	1940	0.64 ± .01	sealed
	1942	0.65 .02	sealed
	1947	0.64 .02	not sealed
counting	1941	0.67 ± .01	not sealed
	1941	0.64 .01	not sealed
	1941	0.68 .01	sealed

On the basis of the measurements of the kimberlite, it might appear that determinations by both methods at the Geophysical Laboratory are slightly too high, but this is not the case when one considers the results from the three laboratories for all the standard rocks. The kimberlite has a very convenient radium content but is not to be highly recommended as the most suitable of the standards because of a tendency to spatter in the fusion furnace whether pre-heated or not. Had no allowance been made for the amount of spattered material, the values in Table I would have been in good agreement with the value assigned by the National Bureau of Standards. It can be stated with assurance that the radium contents of the sediments reported here are accurate to 10 per cent on an absolute basis.

In the analysis of the radioactive data for the purpose of finding the time-depth relations, it is important in some cases to express the radium content on a volume basis. This can be done with a knowledge of the original water content. Those samples in which water was to be determined were weighed within thirty minutes after opening a core liner and later heated at 110 to 120°C and re-weighed. It appears from the results in Table II that the ends of the core liner were not always well sealed. However, there was little evidence of shrinkage, indicating appreciable loss of water, except at the ends of one or two cores.

TABLE II

The Radium Content of Ocean-Bottom Cores;
U.S.C.G.C. *Northwind*, U.S. Navy Antarctic Expedition 1946-47.

(Collected by Dr. J. L. Hough)

Core N–1, Lat 08°56′S, Long 92°05′W, 2150 fathoms, alternating red clay and globigerina ooze.

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
		(red clay)		
N-1- 0	0.0 to	0.3	54.1 (65.5)[a]	18.95
N-1-10	2.0	2.5	—	19.81
N-1-11	4.0	4.5	—	22.39
N-1-14	12.0	12.5	—	23.07
N-1- 4	15.0	15.5	65.9	29.11
N-1-18	25.5	26.0	—	29.67
N-1- 5	30.0	30.8	63.9	26.36
N-1- 7	192.7	194.2	71.5	0.53
		(mixture — predominantly red clay)		
N-1- 2	6.0	6.5	57.3	20.38
N-1-13	9.0	9.5	—	20.43
N-1-17	22.0	23.0	—	20.38
		(mixture — about equal amounts of each type)		
N-1-12	6.6	7.2	—	16.90
N-1-15	18.0	18.5	—	15.86
		(Globigerina ooze)		
N-1-16	20.5	21.5	—	7.01
N-1-19	28.0	29.0	—	6.45
N-1- 3	32.5	33.0	64.3	6.18
N-1-21	40.5	41.5	—	4.75
N-1-23	59.0	60.0	—	4.71
N-1-24	80.0	81.0	—	2.85
N-1-25	100.0	102.0	—	1.40
N-1- 6	124.5	125.7	51.6	0.57
N-1-27	166.5	167.5	—	0.61

[a] Value corrected for observed shrinkage arising from loss of water while in storage in the Lucite liner.

TABLE II (Continued)

Core N-2, Lat. 32°21′S, Long. 105°55′W. 1980 fathoms.
Red clay

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
N-2- 0	0.0 to	0.3	45.8	9.08
N-2- 1	2.0	2.5	46.4	11.09
N-2- 2	6.0	6.5	47.0	13.19
N-2- 3	11.8	12.5	47.0	12.33
N-2-11	16.0	16.5	—	11.16
N-2- 4	25.0	26.0	50.1	7.30
N-2-14	50.0	51.0	—	2.35
N-2-16	80.0	81.0	—	0.72
N-2- 5	100.0	101.0	43.4	0.76
N-2- 6	137.3	138.3	41.5	0.69

Core N-3, Lat. 68°26′S, Long. 179°35′W. 1800± fathoms.
Glacial marine

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
N-3- 0	0.0 to	1.0	—	3.62
N-3- 1	2.0	2.5	38.6	3.49
N-3- 2	8.0	8.5	44.7	4.30
N-3-10	15.0	16.0	—	5.68
N-3- 3	24.0	24.5	47.0	5.07
N-3-12	33.0	34.0	—	4.85
N-3- 4	50.0	51.0	48.4	3.92
N-3-15	80.0	81.0	—	2.41
N-3- 5	100.0	101.0	47.0	1.77
N-3-16	125.0	126.0	—	0.97
N-3- 6	169.0	170.0	47.5	0.62
N-3- 7	227.0	228.0	41.6	0.62

Core N-4, Lat. 69°12′S, Long. 179°34′E. 2040 fathoms.
Glacial marine

Specimen	Depth from top of the core (cm.)		Radium content in 10^{-12} g. per g
N-4- 0	0.0 to	0.5	2.62
N-4- 3	10.0	10.5	3.37
N-4- 5	20.0	21.0	3.70
N-4- 6	30.0	31.0	4.08
N-4- 7	49.8	51.0	4.26
N-4- 8	80.0	81.0	2.65
N-4- 9	120.0	121.0	3.46
N-4-10	160.0	161.0	2.21
N-4-12	225.0	226.0	1.64
N-4-13	259.5	260.5	1.53

No measurements of water content were made in this core.

TABLE II (Continued)

Core N-5, Lat. 70°17′S, Long. 178°23′W. 1635 fathoms.
Glacial marine

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
N-5- 0	0.0 to	1.0	32.1	1.70
N-5- 1	4.0	4.7	40.2	1.79
N-5- 2	9.0	9.6	46.8	2.13
N-5- 3	16.0	17.0	44.2	2.67
N-5- 4	25.0	26.1	43.4	3.03
N-5- 5	55.0	56.0	43.4	2.58
N-5-16	105.0	106.1	—	0.95
N-5- 6	130.0	131.0	45.7	1.29
N-5-17	165.0	166.1	—	1.12
N-5- 7	200.0	201.1	36.6	1.01
N-5- 8	241.3	242.5	40.2	0.77

Core N-10, Lat. 44°14′S, Long. 175°15′E. 300 fathoms.
Greenish gray mud

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
N-10-0	0.0 to	1.0	28.9	1.17
N-10-4	50.0	51.0	35.0	1.14
N-10-6	100.0	101.0	32.5	1.18
N-10-7	140.0	141.0	33.3	1.11
N-10-8	175.0	176.0	29.3	1.18

RESULTS

All determinations of the radium content of the cores from the southern hemisphere are to be found in Table II. The variation of the radium content with the depth in each core is shown in figures 2, 3, and 4. The number of determinations necessary to define a curve varies appreciably for several reasons. In general, the samples must be more closely spaced near the top where the radium content is changing most rapidly, but extreme variations in the type of sediment and sudden changes in the rate of deposition will also require a more detailed survey than would otherwise be the case.

With the exception of the inshore deposit (core N-10), the maximum radium content in all deep-sea cores is to be found at some distance below the surface of the ocean bottom. One can gain a general idea of the recent rate of deposition from the depth at which such maxima occur. The time required

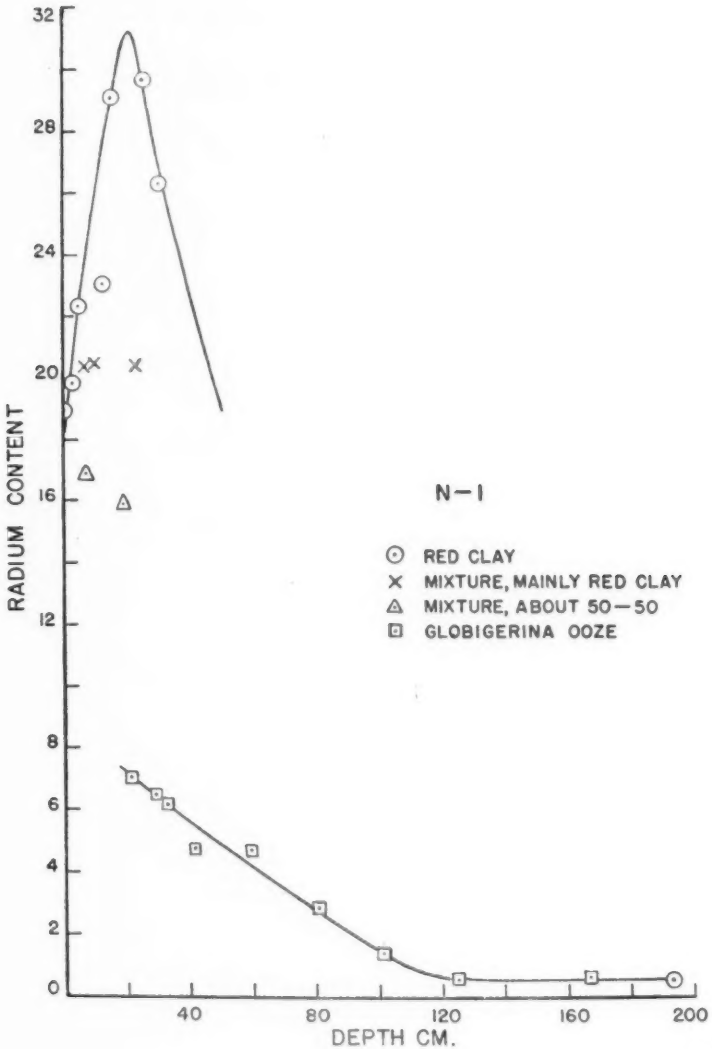

Figure 2. The variation of radium content with depth in a core of alternating red clay and globigerina ooze. The radium content is given in units of 10^{-13} g. radium per g. of sediment dried at 110°C.

to establish the maximum radium content is a function of the initial relation between radium and ionium, but it is not greatly dependent on this parameter and will be between 7000 and 9000 years.[7] A second obvious point on the time scale is where the radium content becomes independent of the depth in the core. This will occur in between 400 and 500 thousand years. Thereafter, this method of dating becomes inapplicable, but reasonable dates can be assigned below such depths if one assumes that the rate of deposition prior to 500 thousand years ago was not greatly different from the rate 400 thousand years ago, and if the distortion in securing the core sample is taken into account.[14]

INDIVIDUAL CORES

Core N-1. (*fig. 2*) Red clay and globigerina ooze have been alternately deposited at this site. It was hoped that two curves could be obtained, one for each type of sediment, and that these curves could be independently analyzed for time. If good agreement resulted, such a procedure would offer strong evidence in support of the basic principles underlying the measurement of time in ocean sediments. Unfortunately, red clay predominates in the top 30 cms., and insufficient globigerina ooze of the required purity is to be found above 20 cms. Below 30 cms., red clay of sufficient purity is unavailable except right at the bottom of the core. Thus, it is only possible to make the desired comparison between the depths of 20 and 30 cms.

It is interesting to note that the radium content, following establishment of complete equilibrium between radium, ionium, and uranium, is independent of the type of sediment. This means that the uranium content, which does not change with time, is the same for both types of sediment in this core and that the striking differences near the top and particularly in the middle are due to very different initial ionium contents.

The maximum recorded radium content in this core occurs at 25.8 cms. (n-1-18). There are very few samples of red clay that show as high a radium content as 29.67×10^{-12} g.

[14] Piggot, C. S., 1941. Factors Involved in Submarine Core Sampling. Bull. Geol. Soc. Amer. 52, 1513.

per g. and there is not one from the "Carnegie" collection of grab samples.[15]

This core sample must be representative of the sediment of more than half of all Pleistocene time at this location.

Core N-2. (fig. 3) This core sample, according to color and texture, is a red clay of very uniform character from top to

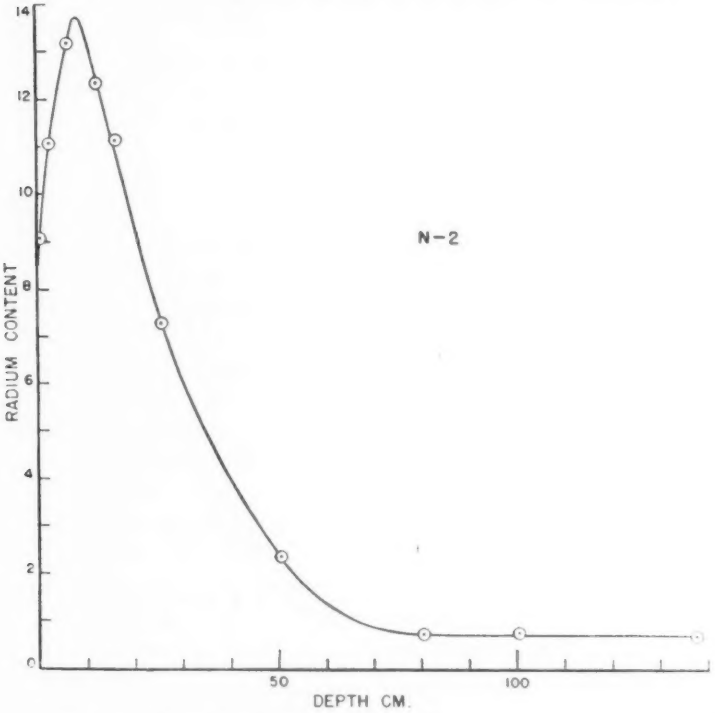

Figure 3. The variation of radium content with depth in a core of red clay. The radium content is given in units of 10^{-12} g. radium per g. of sediment dried at 110°C.

bottom. Consequently, it is to be expected that the variation of the radium content can be expressed by a very smooth

[15] With the aid of figures 2 and 3 it can readily be realized that the radium content of a grab sample has little significance because it depends upon a number of arbitrary factors—the depth of penetration of the grab, and, therefore, the kind of grab employed; whether or not the sample was homogenized; and, if not, the location of the sub-sample for the radium measurement.

curve. This curve exhibits the maximum below the top characteristic of all the other cores. Prior to these determinations, measurements in red clay had been made for only one core of any significant length.[5] The samples for these earlier determinations were undesirably long; and, hence, it was difficult to establish the shape of the curve near the top. For various reasons it was concluded that there was no maximum radium content below the surface; and, thus, the single curve for red clay constituted an exception to the general rule. In view of the curve obtained for the present core of red clay and for the red-clay portion of core N-1 in figure 2, this previous exception may be in doubt.

Core N-2 must be representative of the sediment of most of Pleistocene time.

Cores N-3, N-4, and N-5, (fig. 4) These cores were all secured from the relatively small area lying between Scott Island Bank and Iselin Bank in the Ross Sea, as shown in figure 1. Upon superficial examination they all look quite similar, but at certain depths the sediment is completely free of pebbles and small angular fragments and at other depths such glacial marine debris is abundant. The sediments of all three are tentatively classified as glacial marine. Bearing in mind the two particularly significant positions on the time scale—the maximum value of the radium content and the position where the radium content reaches an essentially constant value—a glance at figure 4 shows that the rate of deposition of the sediment was very different at the site of each one of these cores. It would be interesting to look for the reason for this behavior in continuous ocean-bottom profiles of the area if these were available. The sediment is deposited most rapidly at the site of core N-4 in a small trough south of Scott Island Bank. Events that are characteristic of a given time should have left their records at very different depths in these three cores.

A pronounced dip in the curve is noticeable at the bottom of core N-5. This is to be expected[3] in all cases where the sediment at the bottom is less than 400 thousand years old because of the distortion occurring when the sediment is cored.[14] The distortion causes a shortening of the core sample compared to the original sediment, the amount of the short-

ening being negligible near the top but increasing rapidly near the bottom. A similar effect is present in the curve for core N-4, but it is less noticeable in this region of the curve. Naturally, no such dip is to be seen when the constant equilibrium value of the radium content is attained at some distance above the bottom of a core. The distortion will have no influence on the dating of a specific event, but it is a factor of no slight significance when rates of deposition are to be determined.[6]

One point for each of the cores N-4 and N-5 lies considerably below the curve. Insofar as was possible, pebbles and fragments of rock were removed from the sample whenever they were found in the course of reducing the sample to powder. Such material was considered foreign to the sediment carrying the pertinent radium but some coarse detrital material may have escaped detection. Since such material is almost certainly of much lower radium content than that indicated by the curve, particularly if basic in character, the discrepancy is in the right direction.

Core N-10. (*fig. 4*) As shown in figure 1 this core was secured from the New Zealand plateau in 300 fathoms of water. There is no consistent trend in the radium content, in fact the radium content is most remarkably uniform. It is very doubtful that five samples spaced in like manner over 176 cms. in any non-marine formation would exhibit such a constant radium content.

One might assume that the rate of deposition on the shelf off New Zealand is extremely rapid. The expected increase in radium content might then be hidden in the errors of the measurements, but the age of the sediment at the bottom of core N-10 could hardly exceed 500 years if we employ the greatest spread in the results given in Table II. With due consideration for the distortion in the core sample this would imply a rate of deposition of at least 5 meters per thousand years.

It appears more likely that the sediments of the continental shelf consist largely of detrital material and that the operation of the mechanism that destroys the state of equilibrium between radium and ionium in deep-sea deposition is con-

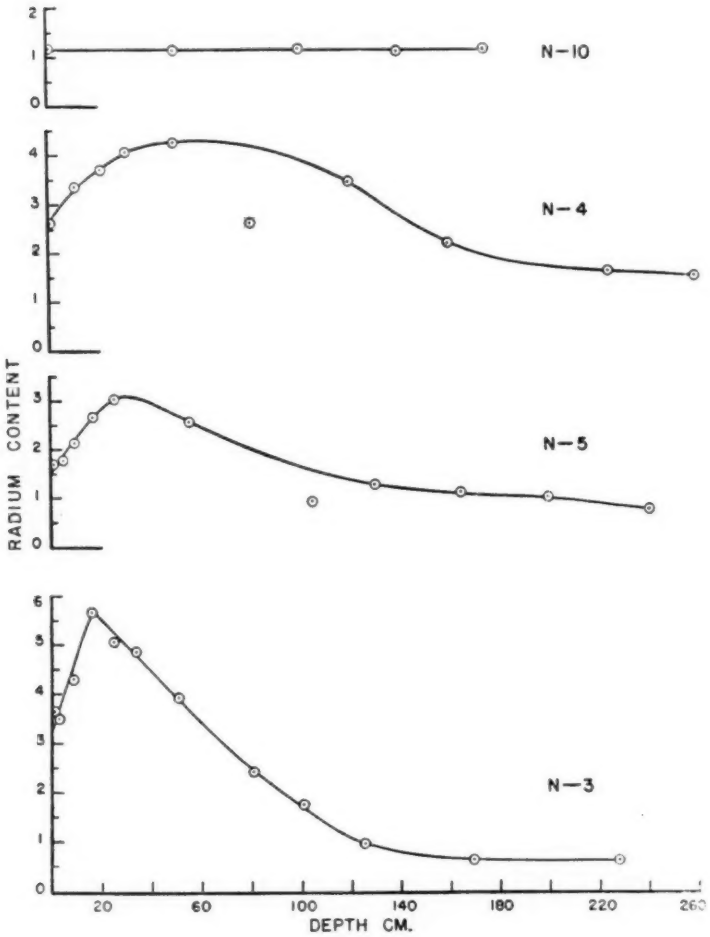

Figure 4. N-3, N-4, N-5 — The variation of radium content with depth in cores of glacial marine deposits of the Ross Sea, arranged in order of increasing rapidity of deposition reading upwards. N-10 — The variation of radium content with depth in a core of greenish gray mud from the New Zealand Plateau. The radium content is given in units of 10^{-12} g. radium per g. of sediment dried at 110°C.

fined to the deep-ocean basins.[16] If this is true, we are deprived of our method of measuring time on the continental shelf, but on the other hand we have a valuable clue to the nature of the mechanism by which the radioactive equilibrium is disturbed in that it requires deep water and the presence of only minor amounts of detrital material for its operation. From this, it follows that the deep-ocean sediments consist almost entirely of chemical and biochemical products. Such a conclusion seems obvious with respect to the calcium carbonate portion of a globigerina ooze, or to the silica of the diatoms or radialaria in these types of ooze; but there has been difference of opinion as to the formation of deep-sea clays devoid of organic skeletons, and as to the nature of the material that binds the organic skeletons together in those oozes consisting largely of organic remains.

ANALYSIS OF RESULTS

The purpose of this investigation was threefold.

Firstly, the results are being analyzed to determine the dates of important events in the Earth's recent history in the southern hemisphere. These events should be indicated by studies, now in progress, of the lithology and of the foraminifera of these cores. It is hoped that at least a tentative answer to the question of contemporaneous glaciation in the two hemispheres can be given by comparing this history with that derived from an earlier and similar co-operative endeavor on cores from the northern hemisphere.

Secondly, the rates of deposition of the sediments at the location of the cores from the southern hemisphere are in the course of evaluation both with respect to the type of sediment and to time in the past.

Thirdly, another analysis of the data is in progress which will add to the available information on the relations between the radioactive elements present in sea water and in ocean bottom sediments. A study of the initial concentrations of the radioelements in the surface of ocean sediments at the location of fourteen cores, including those described here, has revealed

[16] It is not possible to state that the mechanism causing a deficiency of uranium with respect to radium and ionium is also inoperative on the shelf. In order to demonstrate this it would be necessary to examine a very much greater depth of shelf-sediment than is represented by core N-10.

interesting correlations between the concentrations of the radioelements, the rate of their deposition, as distinct from the rate of deposition of the sediment itself, and the fineness of the sediment.

SUMMARY

Seventy determinations of the radium content of ocean sediments are reported here for a number of depths in each of six cores secured from the southern hemisphere. Curves of the radium content as a function of depth below the ocean bottom are similar in all qualitative respects to the curves previously obtained for cores from the North Atlantic and Caribbean Sea.

ACKNOWLEDGMENTS

It is a pleasure to acknowledge the cooperation of the Hydrographic Office of the U.S. Navy in placing the necessary material at the author's disposal through the good offices of Dr. R. H. Fleming. Dr. J. L. Hough contributed greatly to the success of the investigation, primarily in securing the core samples with the aid of his shipmates on the U.S.C.G.C. *Northwind*, and later by collaborating with the author in sampling the cores and assembling the necessary correlative data. The author herewith expresses his sincere thanks to Dr. Hough.

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DOLICHOHARPES AND THE ORIGIN OF THE HARPID FRINGE

HARRY B. WHITTINGTON

ABSTRACT. A new genus of harpid trilobite *Dolichoharpes* (genotype *Eoharpes uniserialis* Raymond, 1925, from the Kimmswick Limestone of Missouri) is diagnosed and the type redescribed. A second species, *D. reticulata*, from the Edinburg Limestone of Virginia, is also described. Specimens of the latter are silicified, and show the morphology of the cephalon in detail. The pits in the bilaminar fringe have flattened bases, but there is no evidence of an opening in the base, and the fringe was therefore apparently not perforate. The glabella and cheek lobes are ornamented by a prominent network of raised ridges surrounding depressions with flat bases, and this ornament merges into the upper lamella of the fringe.

In 1921, Richter suggested that the harpid fringe may have originated from funnel-shaped depressions originally present on the cephalic margin and the doublure. The depressions would be opposed to each other, and by deepening towards each other until the bases came into contact would develop into the characteristic pitted bilaminar fringe. The reticulate ornament of *Dolichoharpes*, if originally present on both the doublure and the dorsal surface, would form an ideal starting point for such a development. The structure of the cephalon strongly suggests that this was the case, and thus affords remarkable support for Richter's hypothesis.

INTRODUCTION

DURING the summer of 1947 the writer spent three months at the U. S. National Museum studying the collection of silicified trilobites from the Edinburg Limestone of Virginia. This visit was made possible by a grant from the Geological Society of America to Dr. G. Arthur Cooper. It is a pleasure to record my indebtedness to Doctor Cooper and to officials of the Society. Through the kindness of Dr. A. S. Romer the types of the trilobites from the Appalachians described by Dr. P. E. Raymond in 1925 were loaned to me for study and comparison with the silicified material. Among these specimens were the cotypes of *Eoharpes uniserialis* Raymond, 1925. My studies of European harpid trilobites showed me that Raymond's species was the representative of a new genus. Fragments of harpids occur very rarely in the silicified Edinburg material, and represent a different species of the same genus. From a block of the limestone, prepared in England, the incomplete cephalon, here described as the holotype of this new species, was obtained. Although this cephalon is incomplete and broken into four fragments, it is large compared with the majority of specimens extracted from this limestone.

The excellent preservation of the silicified cephalon reveals the details of the morphology. It is here suggested that these structures appear to confirm an hypothesis propounded in 1921 by Dr. R. Richter to explain the origin of the harpid fringe. In the following pages the new genus is diagnosed, the two species described, and the structure and origin of the fringe discussed. Reference is also made to the occurrence of a species of harpid in Northern Ireland, believed to belong to the new genus. This species, described long ago by J. E. Portlock, will be redescribed in a forthcoming monograph on British Harpidae.

Dr. G. Winston Sinclair is now studying the American Harpidae, and I am indebted to him for much helpful discussion and for his generosity in suggesting that I should publish this paper in advance of his work. My thanks are also due to the Research Committee, University of Birmingham, for grants in support of my work, and to Mr. L. W. Vaughan of the Geology Department for the preparation of the enlargements and prints from my negatives.

SYSTEMATIC DESCRIPTIONS

Family Harpidae Hawle and Corda, 1847

Genus *Dolichoharpes* Whittington, gen. nov.

Genotype: *Eoharpes uniserialis* Raymond, 1925

Diagnosis: Cephalon strongly convex and ovate in outline, the long axis being longitudinal. Glabella strongly convex, standing well above cheek lobes, highest point at mid-line of posterior margin, first[1] (basal) glabellar lobes large, triangular, first furrows deep, straight, second and third furrows very short and shallow. First glabellar lobes crossed by a prominent ridge running in a curve convex forwards. Between first and second furrows are less strongly ornamented oval areas demarcated by shallow furrows which curve round forwards and outwards from the inner ends of the first glabellar furrows. Eye tubercles and eye ridges well marked. Alae large, gently convex, depressed below level of adjacent parts of cheek lobes. Fringe with deep girder extending to tips of prolongation;[2] internal portion, notably cheek-roll prolongations, broad and

[1] Numbering lobes and furrows forward from the posterior margin of the glabella.

[2] For terminology see Whittington "Redescription of *Eoharpes* Raymond, 1905," Geol. Soc., London, Quart. Jour., **104**, 221-8, pls. xi, xii, 1948.

vertical, external portion narrow with flat lower lamella and concave upper lamella. Marginal band broad. Pits coarse, irregularly and closely spaced in a reticulate pattern except in the upper lamella externally, where there is a tendency to an arrangement of the pits in radiating furrows separated by ridges. Coarse reticulate ornament on cheek lobes and glabella.

Thorax and pygidium unknown.

Stratigraphical Range and Distribution: Middle Ordovician of North America and Ireland. In this paper species of *Dolichoharpes* from the Kimmswick Limestone of Missouri and the Edinburg Limestone of Virginia are described. Material that I have examined in the U. S. National Museum suggests that the genus is also present in the Bromide of Oklahoma. *Dolichoharpes doranni* (Portlock), 1843, from the Bardahessiagh Beds of Pomeroy, Northern Ireland, seems also to belong to this genus. It occurs very rarely in beds which are most probably Middle Ordovician in age (cf. Stubblefield, 1939, p. 60).

Dolichoharpes uniserialis (Raymond), 1925

Plate 1, Figs. 1, 3, 4, 5

[3]1925 Raymond, P. E., p. 15, Pl. 1, figs. 8, 9

Lectotype (here selected). MCZ (Museum of Comparative Zoölogy, Harvard College) 1681, an incomplete external mold of the cephalon.

Other material: MCZ 1682, the second of Raymond's original cotypes, an incomplete cephalon.

Description: The outline of the convex cephalon is oval, the greatest width being at about half the length. The maximum height (at the mid-line of the occipital ring) is about one-third of the width. The length of the prolongations is twice the axial length of the cephalon, and the tips of the prolongations curve inwards. The glabella tapers forwards to the rounded anterior end, the width being greater than the length. The length of the glabella is about one-quarter the length of the cephalon. The deep first glabellar furrows commence

[3] *D. uniserialis* was also described by Bradley (1930, Contr. Walker Mus., 2, No. 6, 244-246, pl. 28, fig. 8-10) but I have not seen this description or the material upon which it was based.

at about two-thirds the length and run straight inwards and backwards to isolate the large triangular first lobes. Shallow furrows connect the inner ends of the first glabellar furrows with the occipital furrow. From the adjacent median portion of the glabella a ridge runs across these connecting furrows and continues across the first glabellar lobes in a curving path, convex forwards (pl. 1, fig. 1). The second glabellar furrows are shallow and short, situated at a point less than half the length of the glabella. From the inner end of the first glabellar furrows shallow furrows run forwards and outwards in a spiral path across the second lobes, finally curving round to run backwards and die out without reaching the first furrows at their outer ends. These furrows thus delimit oval areas which are part, but not all, of the second glabellar lobes. The third glabellar furrows are very short and shallow. The alae are semi-circular in outline, the length equal to about half the length of the glabella. They are gently convex, the outer margins sharply depressed below the adjacent cheek lobes, inwards rising to meet the shallow axial furrows. The occipital furrow is deep, curves forward to the mid-line, and is constricted behind the first lobes. The occipital ring is longest in the median line. The cheek lobes are gently convex in front of the alae, and steeply, almost vertically, bent down at the margins. The eye tubercles, poorly preserved in the type material, are situated opposite a point about one-third the length of the glabella and on the antero-lateral portions of the cheek lobes. The eye ridges run directly inwards to meet the axial furrows opposite the third glabellar furrows. The posterior marginal furrow is shallow, and merges into the alae. The posterior border is increasingly strongly raised laterally, becoming wall-like and continuous with the internal rim.

The fringe is divided by the girder into the broad vertical internal portion and the narrower gently sloping external portion, the upper lamella of which is concave upwards, the external and internal rims are thick, and the broad marginal band has a median ridge, along which the suture presumably runs. MCZ 1682 (pl. 1, figs. 4, 5) has the upper lamella stripped off anteriorly and laterally, and shows that the inner margin of the lower lamella runs in a curved course around the postero-lateral margin of the cheek lobes and thence forwards and slightly upwards to pass in front of the glabella along the

anterior margin of the pre-glabellar furrow. In the internal portion of the fringe the pits anteriorly and laterally are coarse and arranged in a reticulate pattern as far back as about one-third the length of the prolongations, while behind this point they become finer. In the concave depressed portion of the upper lamella the pits appear smaller and, especially laterally, are arranged in radiating rows of 2-3 pits, placed between broad radiating ridges. At the external margin pits appear on the radiating ridges and break up this arrangement. "Along the deepest part of the brim" there is not, as Raymond states a "single line of large pits."

The glabella and cheek lobes are ornamented with a coarse network of raised ridges. Raymond is not correct in stating that the glabella is smooth, and indeed he illustrates the ornament on his pl. 1, fig. 9. At the outer margins of the cheek lobes the ridges of the ornament merge into those between the pits of the upper lamella of the fringe, the boundary between the cheek lobes and the fringe being revealed by the greater relief in the ornament of the latter.

Geological Horizon and Locality: Kimmswick Limestone of Mincke, St. Louis County, Missouri.

Dolichoharpes reticulata Whittington, gen. et sp. nov.

Plate 1, figs. 2, 6, 7; Plate 2

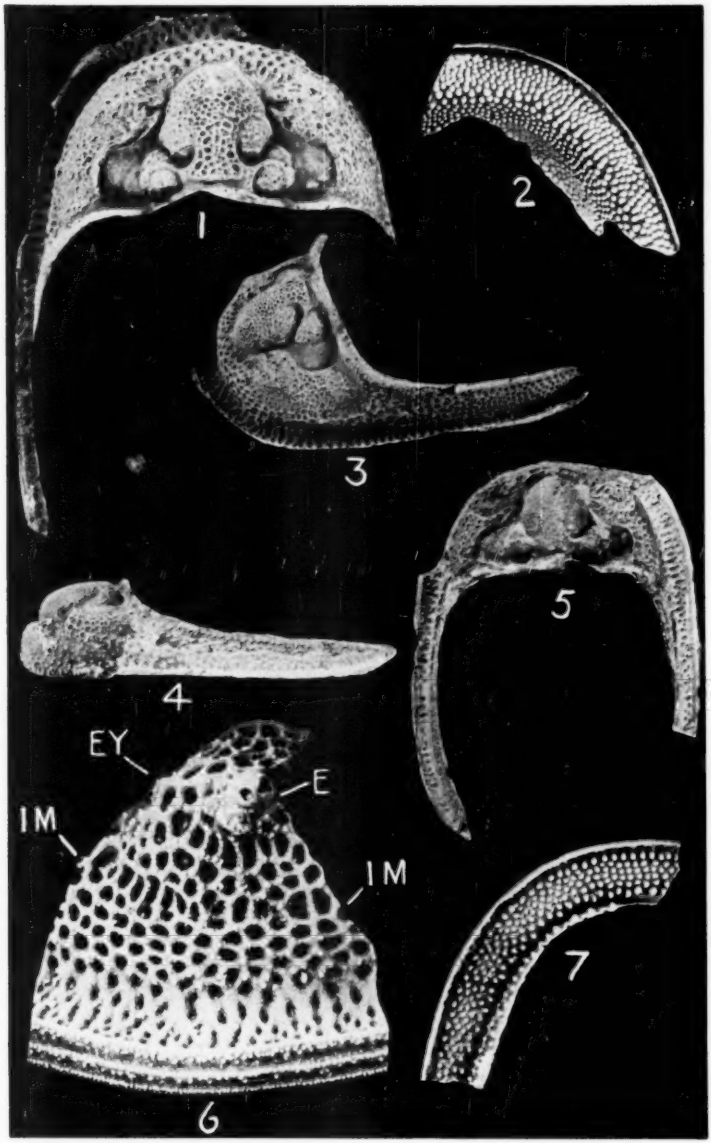
Holotype: The specimen illustrated in plate 1, fig. 6 and plate 2, in the U. S. National Museum.

EXPLANATION OF PLATE 1

Untouched photographs

Figs. 1, 3, 4, 5. *Dolichoharpes uniserialis* (Raymond). Kimmswick Limestone of Missouri. 1, 3. Dorsal and antero-lateral views of a rubber cast of the lectotype, MCZ 1681. Original of Raymond's Pl. I, fig. 9. x 3, x 2. 4, 5. Lateral and dorsal views of a second specimen, MCZ 1682, original of Raymond's pl. I, fig. 8. x 3.

Figs. 2, 6, 7. *Dolichoharpes reticulata* Whittington, gen. et sp. nov. Silicified specimens from Edinburg Limestone of Virginia. 2. Ventral view of left antero-lateral portion of cephalon, showing inner surface of upper lamella and adjacent portion of cheek lobe. x 3. 6. Left antero-lateral portion of holotype cephalon, to show the eye tubercle (E), the eye ridge (EY), and the ornament. IM is the position of the inner margin of the lower lamella, and thus indicates the boundary between the cheek lobe and the fringe. x 6. 7. Dorsal view of inner surface of left antero-lateral portion of lower lamella. The inner edge is broken and the inner margin is not shown. x 4½.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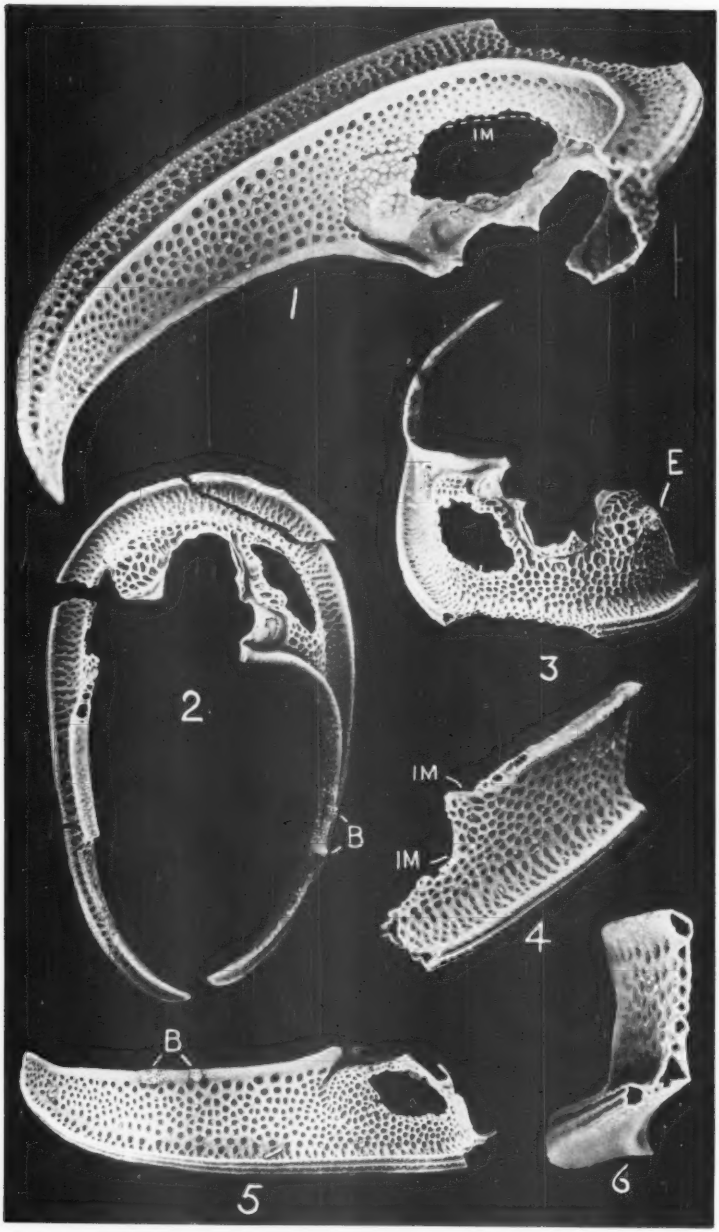
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Other material: two incomplete cephala and various fragments of the fringe, also in the U. S. National Museum.

Description: The glabella of *Dolichoharpes reticulata* is not preserved in the holotype or in any other specimen so far obtained. This species, however, resembles *D. uniserialis* in all other features of the cephalon—outline and shape, the form of the alae, the structure of the fringe, the ornament of the cheek lobes, the position of the eye tubercles—and it is therefore placed in the same genus.

Dolichoharpes reticulata may be distinguished from *D. uniserialis* by differences in detail between the fringes of the two species. In *D. uniserialis* the internal portion of the fringe is inflated anteriorly, and descends vertically to the girder (pl. 1, figs. 1, 4). In *D. reticulata* the corresponding portion of the fringe is not inflated, and slopes steeply down to the girder but is not vertical (pl. 2, fig. 5). The internal rim in *D. reticulata* is very prominent and extends horizontally back from the posterior margin (pl. 2, fig. 5). In *D. uniserialis* the internal rim appears less prominent and slopes gently down and back from the posterior margin (pl. 1, fig. 4). Thus in *D. reticulata* the internal portion of the fringe maintains a greater width posteriorly than in *D. uniserialis*. The external portion of the fringe in *D. reticulata* is wider anteriorly and laterally than in *D. uniserialis*, the pits are more numerous and, while radiating ridges are developed on the upper lamella of *D. reticulata*, they are not so broad as in *D. uniserialis*.

EXPLANATION OF PLATE 2

Untouched photographs

Figs. 1-6. *Dolichoharpes reticulata* Whittington, gen. et sp. nov. Holotype, a silicified specimen broken into four fragments, from the Edinburg Limestone of Virginia. 1. Ventral view of largest fragment, showing lower lamella of fringe with girder. The line IM indicates the inner margin of the lower lamella. Within this line may be seen the ventral surface of the right cheek lobe and ala. x 3. 2. Dorsal view of holotype, the four fragments arranged in approximate relative positions. The larger fragment is shown in figs. 1, 3, 5. B indicates a bryozoan growing on the internal rim. x 2. 3, 5. Anterior and lateral views of the largest fragment of the holotype. E on fig. 3 is the eye tubercle shown on pl. I, fig. 6. B is a bryozoan growing on the internal rim. x 2. 4, 6. Antero-lateral and posterior views of the larger (anterior) fragment of the left prolongation of the holotype, to show the bilaminar fringe in section. IM indicates the position of the internal margin of the lower lamella, within which is a small portion of the postero-lateral corner of the left cheek lobe. x 3, x 4½.

The beautifully preserved silicified specimens of *Dolichoharpes reticulata* reveal the structure of the fringe, the ornament, and other details. The eye tubercle (pl. 1, fig. 6 E) is seen projecting upwards and outwards over the steeply-sloping antero-lateral portion of the cheek lobe. The two raised oval areas (the anterior has adventitious material clinging to it, the posterior is partially broken) are identical in appearance with the two lenses shown to exist in *Harpes macrocephalus* (cf. Richter 1921, p. 194-5, pl. 16, fig. 4, pl. 17, fig. 10). A central circular raised area is also present, corresponding with that in *H. macrocephalus* which is said by Richter not to be a lens. The reticulate ornament on the cheek lobes consists of high ridges enclosing flattened areas. Upon the ridges are scattered small tubercles, those at the intersections being larger and more prominent. The reticulate ornament merges into the pitted fringe structure and the boundary between the cheek lobe and the upper lamella is not easily discerned (pl. 1, fig. 6). The ventral view (pl. 2, fig. 1) reveals the inner edge of the lower lamella curving from the internal rim around the postero-lateral corners of the cheek lobes and running anteriorly just outside the pre-glabellar furrow. The girder is deepest anteriorly, and the rims are prominent, the marginal band having the median ridge along which the suture runs. The sections (pl. 2, figs. 4, 6) show how the flat external portion of the lower lamella is combined with the concave upper lamella. The pits narrow inwards and have a flat base, the bases of opposing pits being in juxtaposition, and the suture passing between them.

In complete specimens of the fringe, or in separate, single lamellae, the base of the pits is represented by a thin layer of silica, much thinner than the walls of the pits or the areas between the pits. Only rarely is seen a tiny hole connecting opposing pits, and where this occurs it is apparently due to breakage or abrasion of the specimen. Since the material is so perfectly silicified, and very fine details of the ornament are preserved, it seems reasonable to infer that in the living animal the bases of the pits were covered by a thin layer of chitin. The structure of the fringe is therefore exactly as described by Richter (1921, pp. 184-7, pl. 16, figs. 4, 5) except that there do not seem to be openings connecting opposing pits. Richter's material was of Devonian age, and it seems that

either the structure of the pits in Devonian harpids was different from that of Ordovician ones, or that the thin layer at the base of the pits was not preserved in Richter's specimens.

An ornament of fine tubercles occurs on the inner and outer rims, and the marginal band. Tubercles occur at the junction of the ridges separating the pits on the upper lamella of the cheek roll and the external portion of the lower lamella.

Geological Horizon and Locality: Basal Edinburg Limestone (? Botetourt member), Middle Ordovician. Holotype from section in field on south side of road and 0.2 miles east of Strasburg Junction, Shenandoah County, Virginia, from crags of dark granular and fine grained limestone about 155 feet southeast of the edge of the quarry dump at the top of the field. Other material, including the two fragments shown in pl. 1. figs. 2, 7, from bed 18 in Geologic Section 19 of Cooper and Cooper (1946, p. 94-5).

STRUCTURE AND ORIGIN OF THE HARPID FRINGE

The structure of the fringe, as described above, appears identical with that of Devonian harpids described by Richter (1921) except that there is no opening, but a thin layer of silica at the base of the pits in the silicified specimens. Extremely fine detail is preserved in tiny specimens from these limestones, and the chitin-replacing silica is very fine-grained. I regard it as unlikely that the silica has grown as a thin layer across what in the living animal was a tiny opening not covered by chitin. Only fragments of young cephala have so far been found but in these the silica layer is of a constant thinness over the glabella, cheek lobes and fringe including the pits, with no suggestion of any opening at the base of the pits.

As has long been recognized, the structure of the harpid fringe is remarkably similar to that of the Trinucleidae (see e.g. Richter, 1921; Størmer, 1930; Whittington, 1941 a, b). The following points of difference may be noted: (a) the canals seen in the marginal part of the cephalon of *Tretaspis* and *Trinucleus* (Størmer, 1930, pp. 100-102) have not been observed in the Harpidae; (b) terrace lines have not been observed on the fringe of harpids; (c) in the Trinucleidae it is believed that the fringe is perforated at the base of the pits (Størmer, 1930, p. 102-103; Whittington, 1941b, p. 509).

The evidence of the silicified Trenton *Cryptolithus* is, however, equivocal, partly because the silicification is coarser than that of the Edinburg material. It seems then, that the Harpidae may include genera with perforate and genera with imperforate fringe, and the same may be true of the Trinucleidae. This condition is theoretically possible if the mode of origin of the fringe discussed below be accepted. It would seem, however, that the adaptive value of a perforate and a nonperforate fringe would be different.

The origin of the fringe and the nature of similar structures in crustaceans and in the tortoise beetle (*Cassida*) have been discussed by Richter (1921, p. 188-191). One of the possible modes of origin of the fringe suggested is as follows (p. 189): if funnel-shaped depressions existed in the borders of the cephalon and in the doublure, opposing pairs of funnels might deepen towards each other, the bases eventually coming into juxtaposition, and finally the funnels opening into each other to form an hour-glass shaped connection between the opposing lamellae. The characteristic ornament of the glabella and cheek lobes of harpids is the raised reticulate pattern of ridges surrounding depressions with flattened bases. Such ornament, extending onto the cephalic borders and the doublure, would form an ideal starting-point for the evolutionary trend suggested by Richter. I believe that *Dolichoharpes* affords clear evidence that this is what actually happened in the Harpidae. The reticulate ornament is strongly developed, and as has already been pointed out, it is difficult, on the dorsal surface, to perceive the boundary between the ornamented cheek lobes and the upper lamella of the fringe, and the similarity between the ornament and the pits in the upper lamella is obvious (see pl. 1, fig. 6; pl. 2, figs. 2-5). It seems that in *Dolichoharpes* the bases of the original depressions in the two lamellae forming the fringe have come into contact and been flattened against each other but do not open into each other. Apparently in the geologically younger species *Harpes macrocephalus* this latter stage is reached, and it may be noted that in this species the ornament of the cheek lobes and the glabella is of small scattered tubercles (Richter, 1921, pl. 16, fig. 3a; pl. 17, fig. 10). The purpose of the pitted structure is clearly to stiffen the thin, bilaminar fringe, as Richter has said (1921, p. 209-210). The development of the thickened rims, the girder, and

radial ridges on the upper lamella also promoted this rigidity.

Reference has been made to the similarity in structure of the harpid and trinucleid fringe. Reticulate ornament is equally characteristic of the cheek lobes and glabella of the Trinucleidae (see Størmer, 1930; Whittington, 1941a), and it is possible that the pitted fringe originated in the same way as suggested for the Harpidae. It seems to me that this similarity of structure and origin need not necessarily be regarded as implying that the two families are closely related.

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REVIEWS

The Essentials of Organic Chemistry; by C. D. PORTER and T. D. STEWART. Pp. vi, 394. Boston, 1948 (Ginn and Co., $4.00).—This book is designed to fill the need for a text to be used by pre-medical, pre-dental and pre-nursing students, as well as by students in the liberal arts, in a semester course in introductory organic chemistry. The text, written by the same authors as the earlier and more complete *Organic Chemistry* (Ginn and Co., 1943), presents a survey of the essential chemistry of the aliphatic and aromatic compounds, and special brief chapters are devoted to the carbohydrates, the proteins and amino acids, dyes and indicators, natural and synthetic drugs, the enzymes, vitamins and hormones, and to the sterols. Stereochemistry is introduced at an appropriate location, and theoretical material, such as the nature of the chemical bond and resonance, is reduced to its barest descriptive essentials. The relations and applications of organic chemistry to the biological sciences and to industry are concisely and lucidly illustrated. Each chapter is accompanied by a set of exercises and a set of problems which have been selected with due consideration of the mastery which a student can be expected to attain in such a courrse. It is the opinion of this reviewer that the authors have achieved their task of condensation and simplification with a minimum sacrifice of useful knowledge, and that the text can be used with every expectation of success. ALFRED H. FRYE

Manual of Coastal Delineation from Aerial Photographs; by P. G. MCCURDY. U. S. Navy Department, Hydrographic Office Publication No. 592, Pp. iii, 143; 202 figs. Washington, D. C., 1947 (The Hydrographic Office, Washington 25, D. C., $1.50).—This manual is intended primarily, as the author states, to make photogrammetrists, hydrographic engineers, and cartographers conscious of each other's problems, to help those delineating coastal areas from aerial photographs, and to help those using the delineated photographs in compiling coastal charts. The manual, nevertheless, has definite geological appeal, and is very useful in geologic instruction. It is well organized, the diagrams and photographs appear on pages opposite the discussions to which they pertain, and reproduction of the photographs is excellent. The text is typed with right-hand margin control, and both text and photographs are offset reproduced. Of the 202 figures, 162 are aerial photographs (97 vertical stereopairs and 59 obliques); 25 are diagrams and 15 are chartlets (small nautical charts with soundings shown in fathoms). Many of the instructive diagrams are

taken from Johnson's *Shore Processes and Shoreline Development*, 1938, and Lobeck's *Geomorphology*, 1939.

The four major coastal types, explained in detail with accompanying diagrams, chartlets, and oblique photographs, are those proposed by Johnson in 1938. Two other types of coasts considered and pictured are the vegetation-bordered coast (capably described in *Pacific Landforms and Vegetation*, U. S. Navy Photographic Intelligence Center Report No. 7, 1945), and the man made coast. A seventh type, the ice barrier coast, is to be described in a later edition. From the viewpoint of persons concerned with the description of a coast and not too greatly interested in its geologic history, it would seem logical and simpler to use a descriptive classification of coasts, instead of the genetic approach of Johnson, which requires field evidence for final proof. Because of this requirement, some of the coasts described in the manual, based only upon photographic interpretation, are subject to correction. A descriptive classification also would be in keeping with the well-founded vegetaton-bordered coast, the man made coast, and the ice barrier coast.

The chapter "Coastal Items To Be Mapped" is beautifully illustrated with vertical stereopairs. It is actually a chapter in beach interpretation from aerial photographs, and is suitable for teaching geomorphology and photographic interpretation. The chapters on "Manuscript Symbols" and "Recommended Procedure for Delineation of Coastal Features" are for the cartographer and photogrammetrist. The final chapter illustrates the problems in delineation of coastal features by carefully selected and arranged oblique photographs, with duplicate sets of vertical stereopairs of part of the same area shown on the page opposite. The first set of stereopairs is untouched; the duplicate set is marked to show the delineation of the particular coastal features under discussion.

From a geological standpoint it is regrettable that several small errors of usage and misapplication of terms exist. After careful scrutiny, the reader will find that the operational mechanics of longshore currents are not fully understood by the author. As many photographs are of great geologic interest, it is disappointing that their localities are not given. The manual is valuable for teaching the physiography of coasts and shoreline features to elimentary classes in geology and geomorphology, and it serves as a worthwhile supplement to courses in aerial photographic interpretation and photogeology.

SIDNEY E. WHITE

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